THE SPILITE-KERATOPHYRE ASSOCIATION OF WEST TASMANIA AND THE ORE DEPOSITS AT MT. LYELL, ROSEBERY AND HERCULES.

by

MICHAEL SOLOMON B.Sc. (Hons.), M.Sc.

Submitted in fulfillment of the requirements for the degree of Doctor of Philosophy

UNIVERSITY OF TASMANIA

HOBART

1964

This thesis contains no material which has been accepted for the award of any other degree or diploma in any University, and to the best of my knowledge and belief, contains no paraphrase of material previously published or written by another person, except when due reference is made in the text of the thesis.

M. Solomon.

University of Tasmania.

November, 1964.

CONTENTS

VOLUME I

Page No.

Page No.

REFERENCES

APPENDIX A

Copies of papers by the author referred to in the text.

APPENDIX B

Supporting papers.

LIST OF FIGURES

Following Page

LIST OF PLATES

Following Page

LIST OF TABLES

ABSTRACT

Western Tasmania was the site of geosynclinal sedimentation during the Cambrian and the late Precambrian. In the Early Cambrian there was a marked change from Precambrian (?) deposition of sandstones, mudstones and dolomite to a typical synorogenic greywacke sedimentation that continued into the Upper Cambrian. This change was accompanied by extrusion of spilitic and keratophyric volcanics.

The basic volcanics are best exposed on King Island and include pillow lavas, breccias called isolated-pillow, broken-pillow, and broken-flow breccias and also tuffs similar to palagonite tuffs in Iceland and "aquagene tuffs" in British Columbia. The pillows (which develop by a combination of surface tension and gravitational effects) and also the finely globular tuffs, indicate an aqueous environment.

The spilites contain three principal types:

(a) picrite

(b) tholeiitic and

(c) augite-rich.

They are generally low in titania and rich in alumina. Differentiation is characterised by iron enrichment but follows neither alkaline, tholeiitic or calc-alkaline trends.

In some areas the volcanics show considerable alteration (with development of hydrogrossular, chlorite, quartz, sericite and calcite)

most of which is considered to be of deuteric origin.

The spilites are associated with small amounts of keratophyre and quartz keratophyre.

Within the Cambrian geosyncline, around the margin of a geanticlinal ridge in the centre of Tasmania (the Tyennan Geanticline) there developed a volcanic arc (the Mt. Read Volcanic Arc) which commenced to form at much the same time as the spilites but continued to develop probably well up into the Cambrian.

This arc is composed largely of quartz keratophyres, both sodic and potassic, which form massive flows and probably also intrusive bodies. Albiteandesites occur near Queenstown. Basal members are characteristically potassic and those near Rosebery (the Primrose Pyroclastics) show many features of ignimbrites or froth-flows.

Lenses of banded tuff and mudstone in the Arc indicate aqueous deposition but at least some of the volcanic activity was probably subaerial. Entry of nuées ardentes into the sea, or slumping off the flanks of volcanoes, or submarine volcanic activity, probably gave rise to density currents which swept into the flanking basin to deposit feldspar-rich fragmental rocks.

Variation diagrams show calc-alkaline affinities for the keratophyres.

As in other spilite-keratophyre provinces, the quartz keratophyres are far more abundant than other volcanics and differentiation from spilitic magma seems unlikely.

Rocks associated with the spilites and keratophyres include later serpentinites and gabbros, and sedimentary cherts, forming a typical ophiolite suite. The Mt. Read Volcanics are intruded by small, partly concordant, sodic and potassic granites.

It is tentatively concluded that the spilites and keratophyres are derived from independent magmas intruded during a phase of renewed sagging and stretching of the geosyncline. The intake of water, and possibly other material, into the magmas may have affected the subsequent crystallisation, leading to formation of albite-diopside and albite-chlorite rocks, or the magma may have been hydrous from inception.

Some potassic quartz keratophyres contain thick veins of hematite and magnetite at several points along the Mt. Read Volcanic Arc and several of these occurrences have associated copper mineralization.

Most of Tasmania's metallic ore deposits (mainly tin, lead, zinc and copper) occur in areas of Cambrian sedimentation and particularly where spilites or keratophyres are abundant. Some of the deposits are clearly related to Devonian granites but the two largest deposits, Mt. Lyell (copper) and Rosebery-Hercules (zinc-lead-copper) occur in the Mt. Read Volcanic Arc and lack any obvious granitic source.

Mt. Lyell ores are mainly pyrite-chalcopyrite, and chalcopyrite-bornite-pyrite assemblages with very minor pyrite-sphalerite-galena lodes. They form lenses replacing cleaved and hydrothermally altered Mt. Read Volcanics adjacent to Ordovician Owen Conglomerate. They were deposited after alteration and cleavage development, both related to the Devonian Tabberabberan Orogeny. The hydrothermal alteration resulted in quartz-sericite and quartz-chlorite "schists" with minor pyrophyllite, and probably took place between 350 and 600° at about 800 atmospheres. The temperature of ore deposition is unknown but may have risen to 700° C.

The NW cleavage, and parallel thrust and transcurrent faults, provided local structural controls but more important controls appear to be the Great Lyell Fault Zone, which dates back to the Cambrian, and the distribution of certain potassic keratophyres. Massive hematite bodies at the base of the Owen Conglomerate give a possible indication of pre-Ordovician mineralization.

The Rosebery and Hercules ores form discrete pyrite-sphalerite-galena-chalcopyrite lenses replacing a fine-grained tuff that overlies the Primrose Volcanics. This tuff exerts a pronounced stratigraphic control, and bedding and cleavage are also important on a smaller scale in controlling the orientation of individual lenses. The ores, which are locally banded, follow sericite-carbonate alteration and Devonian cleavage. Local structural controls are not obvious at Rosebery and Hercules but both mines (and other small deposits) lie on a major fault zone (the Rosebery Fault Zone) which probably was active in Cambrian times.

The ores show chemical features typical of magmatic deposits and their final phase of deposition was clearly of Devonian age. However, there is evidence of a pre-Ordovician phase of mineralization and the ores may be of volcanic origin, deposited within the Mt. Read Volcanics soon after their eruption.

ACKNOWLEDGMENTS

Grateful acknowledgment is made of help given by a number of geologists in academic and industrial circles. Particular thanks are due to Professor S.W. Carey and Dr. A. Spry, and also to Dr. R. Green, Mr. R. Ford and Mr. M.R. Banks, all of the Geology Department, University of Tasmania, for many helpful discussions.

The Mt. Lyell Mining and Railway Co. Ltd. (Queenstown) and the Electrolytic Zinc Co. (Rosebery) have been most generous in giving ~~written~~ access to information and the Mt. Lyell Co. has defrayed the cost of two chemical analyses.

Many discussions have been held with Mr. G. F. Hudspeth

(Assistant General Manager) and Mr. R.G. Elms (Chief Geologist) of the Mt. Lyell Co., and Mr. V. Cottle and Mr. R. Pratten of the Electrolytic Zinc Co.'s. geological staff.

Special mention should be made of the stimulus provided by Dr. B. Campana and the late Mr. M.L. Wade (of Conzinc Rio Tinto of Australia, Ltd.) during the early stages of the field work.

The writer has benefited from his association with several of his Honours students, particularly Mr. C.C. Brooks, Mr. D.I. Groves and Mr. G.L. Hills.

Mr. H. Bartlett, of King Island, gave invaluable assistance during the field work on King Island.

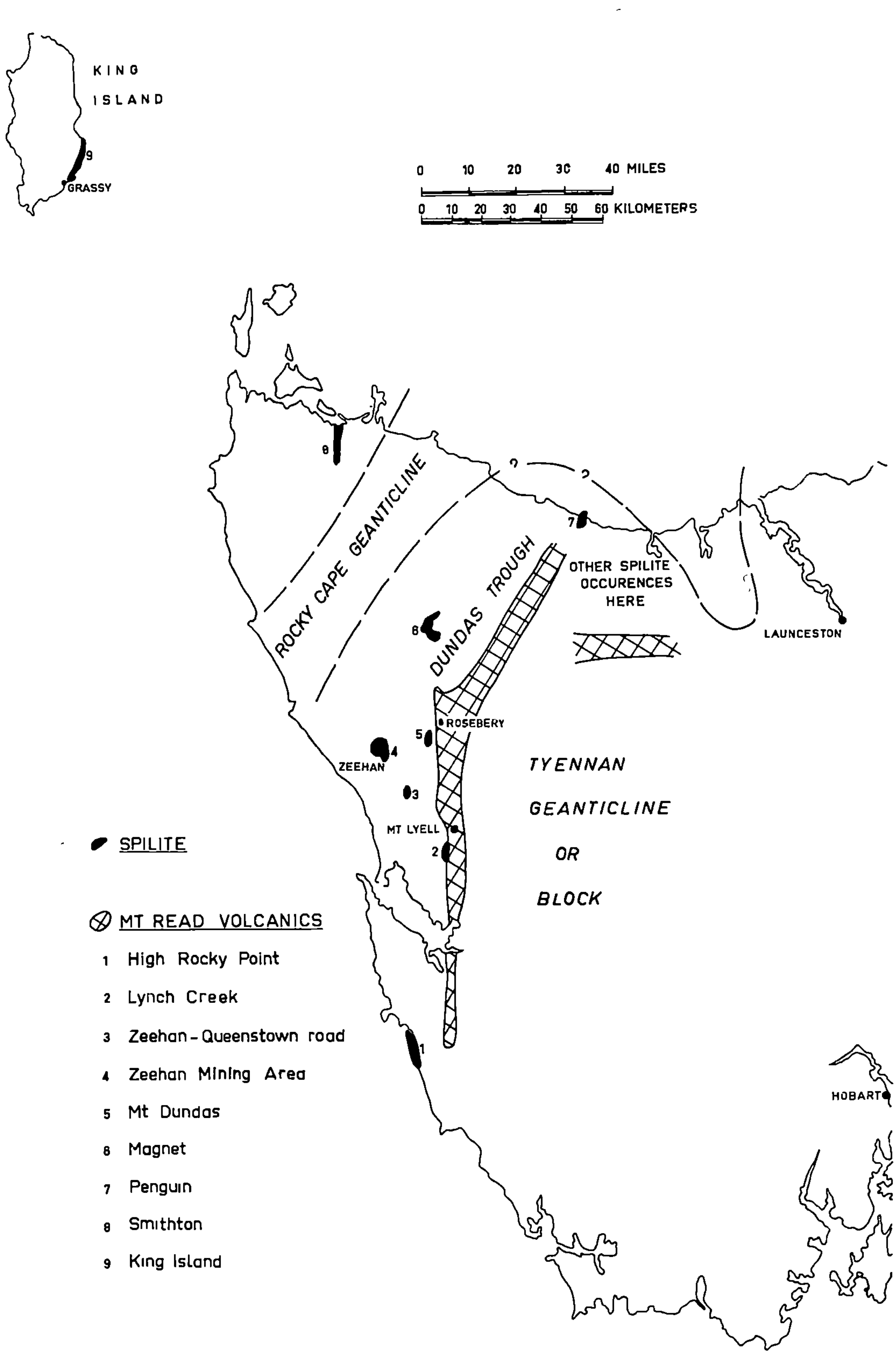
Finally, the writer acknowledges financial assistance received from research grants provided by the University of Tasmania.

INTRODUCTION

The origin of the Rosebery, Mt. Lyell and associated ore deposits of the West Coast Range, Western Tasmania, has remained uncertain since their discovery in the late 19th century, though most workershave called on the aid of an unseen granitic source. The relatively recent world-wide recognition of the possible importance of vulcanism in ore deposition focuss ed the attention of thewriter on the spatial connection of the Cambrian volcanics of Western Tasmania and the metalliferous ore deposits, and in particular on the occurrence of the Rosebery and Mt. Lyell deposits within a prominent volcanic arc (Fig. 1).

As already outlined by Carey (1947a, 1953), Banks (1962a), and Solomon (1962, in press), vulcanism was widespread in the early phases of Cambrian sedimentation in Tasmania. At this time Tasmania formed part of an extensive geosyncline (the Tasman Geosyncline) extending north into Victoria and South Australia, and Cambrian greywackes and mudstones were being deposited over a rather complicated topography. Earlier deposits (supposedly Precambrian) consisted essentially of two groups (see Spry, 1962a); the Younger Precambrian (Quartzites, shales, etc.) and the Older Precambrian (more severely deformed garnet-muscovite schists, sheared quartzites, amphibolites, etc.). Spry believes the Older and Younger rocks are separated by the Frenchman Orogeny and that the Younger were laid down in a miogeosyncline restricted by the growth of a geanticlinal zone (the Tyennan Geanticline, or Tyennan Block of Carey, 1953) in the centre of Tasmania (Fig. 1). After another phase of deformation, the Penguin Orogeny, Cambrian deposition followed but was limited in extent not only by the Tyennan Geanticline but also by another ridge in the north-west of the island, the Rocky Cape Geanticline (Fig. 1). The main centre of Cambrian

Fig. 1: Map of West Tasmania showing the geanticlinal area of Cambrian times and the occurrences of spilitic rocks in relation to the Mt. Read Volcanic Arc.



KING ISLAND
9
GRASSY
0 10 20 30 40 MILES
0 10 20 30 40 50 60 KILOMETERS
8
ROCKY CAPE GEANTICLINE
DUNDAS TROUGH
7
OTHER SPILITE OCCURENCES HERE
LAUNCESTON
6
ROSEBERY
5
4
ZEEHAN
3
MT LYELL
2
TYENNAN GEANTICLINE OR BLOCK
1
HOBART
SPILITE
MT READ VOLCANICS
1 High Rocky Point
2 Lynch Creek
3 Zeehan - Queenstown road
4 Zeehan Mining Area
5 Mt Dundas
6 Magnet
7 Penguin
8 Smithton
9 King Island

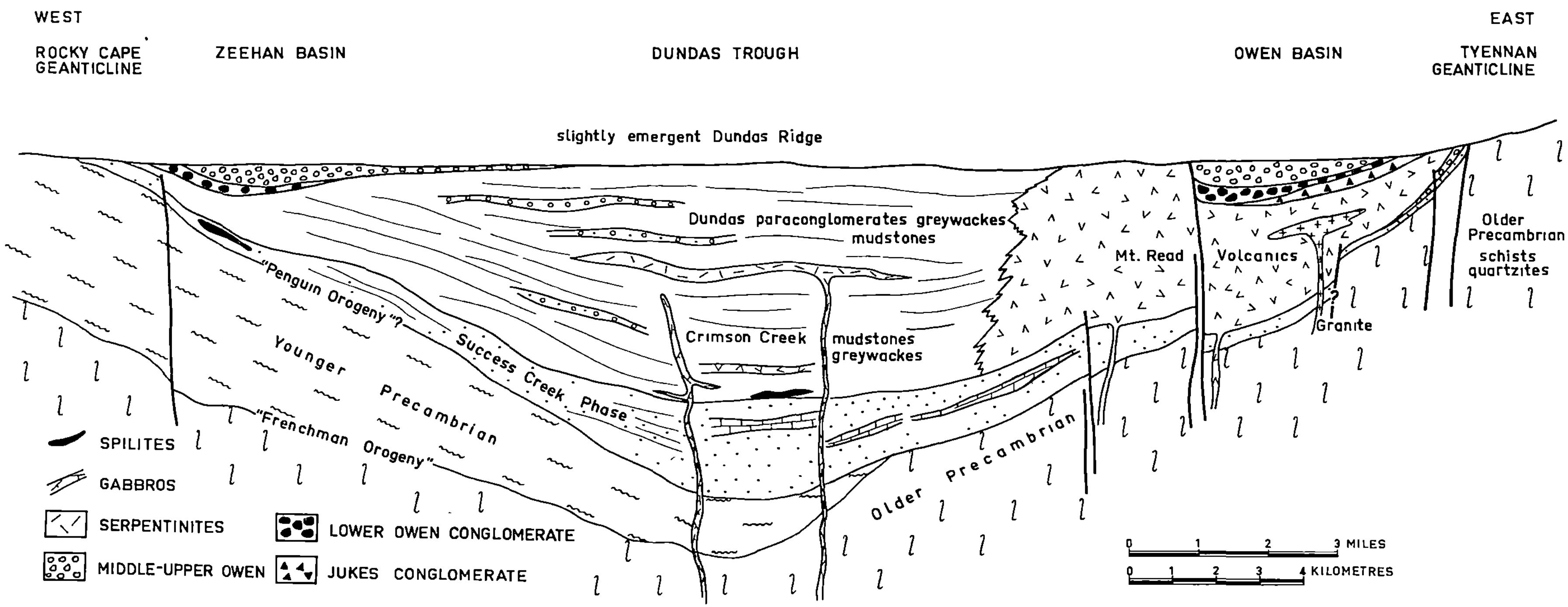
deposition was in the Dundas Trough.

The writer has suggested (in press, Appendix A Paper 1) that the Cambrian greywacke-mudstone successions (the Crimson Creek Argillite and the Dundas Group) were preceded by a dolomite-sandstone sequence (the Success Creek phase) found in many areas conformably underlying the Cambrian successions, and probably of early Cambrian or late Proterozoic age. The stratigraphic position and distribution of the Cambrian and Precambrian rocks are shown somewhat diagrammatically in cross section in Fig. 2.

The Dundas Group (defined by Elliston, 1953, as part of the Dundas Slates of Ward, 1909, but considerably modified by Banks, 1962a, p. 132, and Blissett, 1962, p. 32) is fossiliferous and extends from the Lower Middle Cambrian to the Lower Upper Cambrian; in the Zeehan-Rosebery area at least it overlies the Crimson Creek Argillite (defined by Taylor, 1954) which is unfossiliferous. The two units total some 4.5 to 5.5 km. (Blissett, 1962).

Cambrian vulcanism was characterised by the wide distribution of spilitic flows (Fig. 1) and the growth of a keratophyric arc along the western margin of the then exposed, or shallowly submerged, Tyennan Geanticline. This arc (the Mt. Read Volcanic Arc) became the site of the shallow basin of Lower Ordovician age in which the bulk of the Owen Conglomerate was deposited, following a phase of uplift by faulting and gentle folding (the Jukesian Orogeny).

Fig. 2: Cross-section, slightly generalised, from Zeehan to Mt. Murchison, as at the close of Owen Conglomerate time.


WEST
EAST
ROCKY CAPE GEANTICLINE
ZEEHAN BASIN
DUNDAS TROUGH
OWEN BASIN
TYENNAN GEANTICLINE
slightly emergent Dundas Ridge
Dundas paraconglomerates greywackes mudstones
Mt. Read
Volcanics
Older Precambrian schists quartzites
Granite
"Penguin Orogeny"?
Success Creek Phase
Younger Precambrian
"Frenchman Orogeny"
Crimson Creek
mudstones greywackes
Older Precambrian
SPILITES
GABBROS
SERPENTINITES
MIDDLE-UPPER OWEN
LOWER OWEN CONGLOMERATE
JUKES CONGLOMERATE
0 1 2 3 MILES
0 1 2 3 4 KILOMETRES

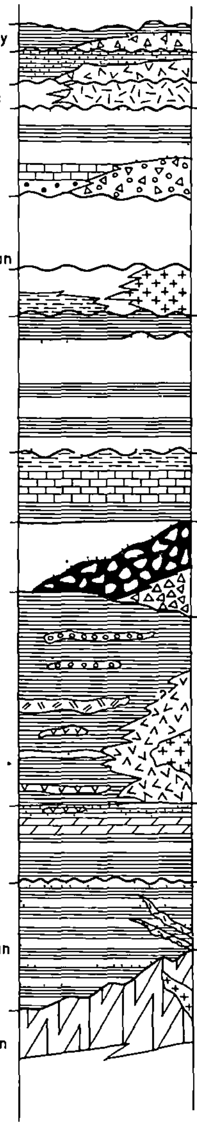
Widespread marine transgression followed Owen deposition and relative stability reigned until the onset of the Tabberabberan Orogeny in the Middle Devonian, a major phase of faulting, folding, granite intrusion and mineralization. Clearly many of Tasmania's mineral deposits are related to this orogeny (Hall and Solomon, 1962). Succeeding phases of western Tasmania's history have involved relatively little sedimentation and essentially epeirogenic activity, and much of the ground has been exposed to erosion during the Carboniferous, possibly throughout the Cretaceous, and since the mid-Tertiary.

The geological history is summarised in Table 1, this being a slightly modified form of previous tables (Solomon, 1962; and Appendix A Paper 1) for the whole State. The reader is referred to Appendix A Paper 1 for a more detailed summary.

This thesis includes a study of the distribution, nature and origin of Cambrian spilites and keratophyres and a study of the nature and genesis of the main ore deposits of the Mt. Read Volcanic Arc, none of which have any obvious connection with granite intrusions.

The spilites south of Penguin (see Fig. 1) were not examined

TABLE 1



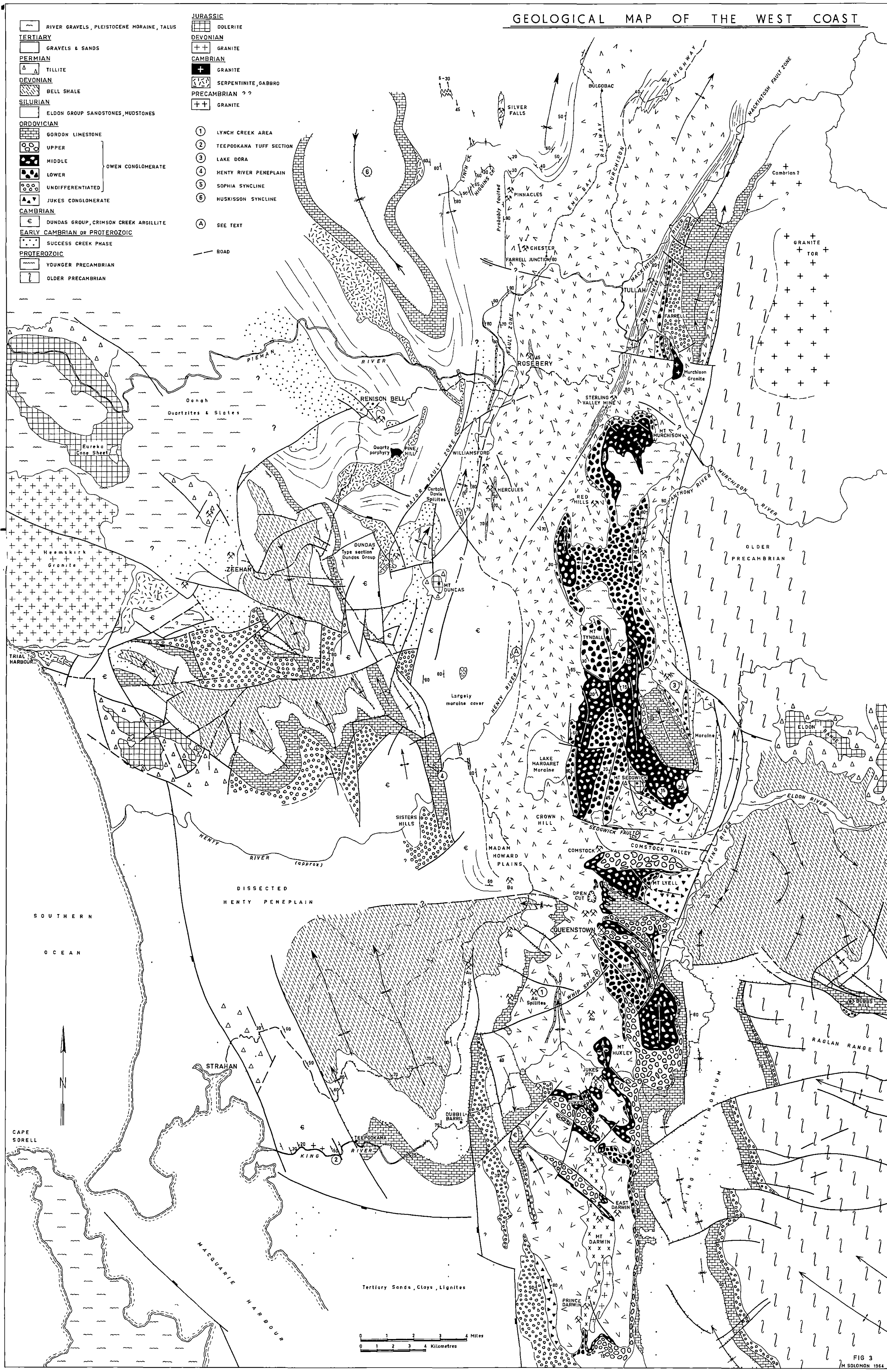
SEDIMENTS	AGE	THICKNESS IN FEET x 10	TECTONIC & IGNEOUS ACTIVITY, MINERALIZATION
Largely non-marine sands	Quaternary		EPEIROGENIC UPLIFT, HORST & GRABEN STRUCTURES
	Tertiary		Basalt flows, volcanic necks, bauxite
	Jurassic		Dolerite sills & cone sheets
Sandstone Conglomerate Tillite Coal measures	Triassic & Permian	5	
	Carboniferous & late Devonian		UPLIFT & EROSION KANIMBLAN OROGENY ? Granite stocks, contact metasomatic ores sulphide-cassiterite "sills", lead-zinc fissures, regeneration (?) of Cu,Pb,Zn ores
Eugenana beds Cave filling	Middle Devonian		TABBERABBERAN OROGENY
Eldon Group Mathinna beds Sandstone Shale	Lower Devonian to Silurian	10	Miogeosyncline, little disturbance
			BENAMBRAN OROGENY
Junee Group Siltstone Gordon Limestone Florentine Valley Mudstone	Ordovician	6	Miogeosyncline
Caroline Ck Sandstone Owen Conglomerate Jukes Breccia	Lower Ordovician to Upper Cambrian	0-3	Partly terrestrial environment
			JUKESIAN OROGENY troughs & rift valleys, tensional faulting
Dundas Group and Crimson Creek Argillite Mudstone Greywacke Paraconglomerate	Upper Cambrian to Middle Cambrian	20	Eugeosyncline ophiolite association Serpentinites, gabbros Cu-Ni, Os Ir mineralisation Murchison & Darwin granites (Jukesian) Mt Read volcanics, iron ores and Cu Pb Zn mineralisation Spilites, tensional faulting (?)
Success Creek Phase Sandstone Dolomite	Early Cambrian	8?	Miogeosyncline
			PENGUIN OROGENY
Rocky Cape, Carbine Grps Oonah Formation Quartzite Shale	Younger Pre-Cambrian	7+	Dolerite & amphibolite dykes magnetite lenses FRENCHMAN OROGENY Granite stock at Granite Tor (?)
Franklin, Mary Grps Schist Quartzite Amphibolite	Older Pre-Cambrian		

and the Mt. Read Volcanic Arc was examined only between a point about 15 miles north-east of Rosebery and about 20 miles south of Mt. Lyell.

Field work occupied about 10 months in the summer seasons of 1957-58 and 1958-59, while the writer was employed by the Rio Tinto Mining Company, and almost five months spread over the period 1960 to 1963 inclusive. Some of the detailed mapping at Mt. Lyell was done during 1955 and 1956 while in the employ of the Mt. Lyell Mining and Railway Co. Ltd. The laboratory work was undertaken in the period from early 1960 to early 1964 at the Department of Geology, University of Tasmania. Mapping at 1:1,200 was carried out at Mt. Lyell and Renison Bell, and mapping at about 1:15,000 over most of the Mt. Lyell-Rosebery section of the volcanic arc and for several miles west of this part of the arc. Small areas such as the Magnet-Waratah area, the Smithton area, and the south-east coastline of King Island were mapped on a similar scale.

A geological map including work by other geologists (particularly of the Tasmanian Department of Mines and the Rio Tinto Mining Co.) is included as Fig. 3.

Fig. 3: Geological map of Western Tasmania. Compiled from work by the writer, from Blissett (1962) for the Zeehan-Dundas area, and from work by Campana, King, Fraser and others of Conzinc Rio Tinto for the Mt. Murchison-Mt. Tyndall area.


GEOLOGICAL MAP OF THE WEST COAST
RIVER GRAVELS, PLEISTOCENE MORAINE, TALUS
TERTIARY
GRAVELS & SANDS
PERMIAN
TILLITE
DEVONIAN
BELL SHALE
SILURIAN
ELDON GROUP SANDSTONES, MUDSTONES
ORDOVICIAN
GORDON LIMESTONE
UPPER
MIDDLE
LOWER
UNDIFFERENTIATED
OWEN CONGLOMERATE
JUKES CONGLOMERATE
CAMBRIAN
DUNDAS GROUP, CRIMSON CREEK ARGILLITE
EARLY CAMBRIAN OR PROTEROZOIC
SUCCESS CREEK PHASE
PROTEROZOIC
YOUNGER PRECAMBRIAN
OLDER PRECAMBRIAN
JURASSIC
DOLERITE
DEVONIAN
GRANITE
CAMBRIAN
GRANITE
SERPENTINITE, GABBRO
PRECAMBRIAN ??
GRANITE
1 LYNCH CREEK AREA
2 TEEPOOKANA TUFF SECTION
3 LAKE DORA
4 HENTY RIVER PENEPLAIN
5 SOPHIA SYNCLINE
6 HUSKISSON SYNCLINE
A SEE TEXT
ROAD
SILVER FALLS
BULGOBAC
MURCHISON HIGHWAY
MACKINTOSH FAULT ZONE
EMU BAY RAILWAY
PINNACLES
CHESTER
FARRELL JUNCTION
Probably faulted
Cambrian ?
GRANITE TOR
TULLAH
MT FARRELL
Murchison Granite
ROSEBERY
PIEMAN RIVER
RENISON BELL
Oonah Quartzites & Slates
Eureka Cone Sheet
STERLING VALLEY MINE
MT MURCHISON
Quartz porphyry
PINE HILL
WILLIAMSFORD
MAJOR FAULT ZONE
ANTHONY RIVER
MURCHISON RIVER
HERCULES
Curtain Davis Spilites
RED HILLS
DUNDAS Type section Dundas Group
Heemskirk Granite
ZEEHAN
MT DUNDAS
OLDER PRECAMBRIAN
MT TYNDALL
TRIAL HARBOUR
Largely moraine cover
HENTY RIVER
Moraine
ELDON RANGE
LAKE MARGARET Moraine
MT SEDGWICK
ELDON RIVER
SISTERS HILLS
CROWN HILL
SEDGWICK FAULT
COMSTOCK
COMSTOCK VALLEY
KING RIVER
MADAM HOWARD PLAINS
HENTY RIVER (approx)
MT LYELL
OPEN CUT
DISSECTED HENTY PENEPLAIN
SOUTHERN OCEAN
QUEENSTOWN
MT OWEN
WHIP SPUR
Au Spilites
BUBBS HILL
RAGLAN RANGE
MT HUXLEY
STRAHAN
JUKES PTY
MT JUKES
KING SYNCLINORIUM
DUBBIL BARRIL
CAPE SORELL
TEEPOOKANA
KING RIVER
East DARWIN
MT DARWIN
MACQUARIE HARBOUR
Tertiary Sands, Clays, Lignites
PRINCE DARWIN
0 1 2 3 4 Miles
0 1 2 3 4 Kilometres
M SOLOMON 1964
FIG 3

SPILITES

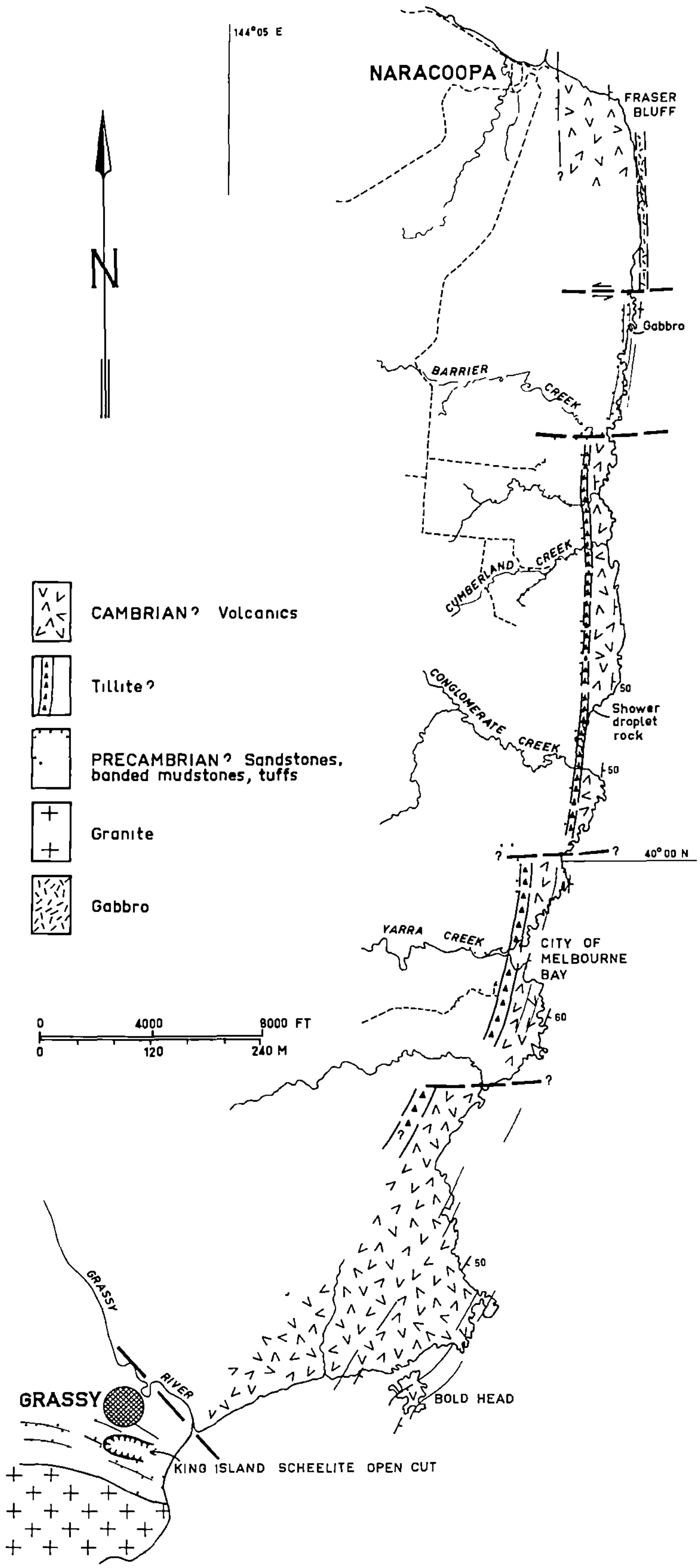
Principal Rock Types and their Mode of Eruption

Spilites, albite basalts and dolerites occur on the north and west coasts of Tasmania and at King Island (Fig. 1). The excellent exposures at King Island have been described briefly by Scott (1951a) and she has also described some of the other occurrences on the West Coast (e.g. Smithton; Lynch Creek, Queenstown; and Penguin). Well-exposed areas like King Island and the Wanderer River district reveal flows from a few metres to over 30 metres thick interbedded with minor tuff and siltstone beds; in other areas (Lynch Creek, Zeehan, etc.) the bedding is discernible only with difficulty and individual flows cannot be isolated. Though most workers have accepted a Cambrian age for the volcanics (on lithology and structural interpretations), they could be Proterozoic. The underlying sediments are everywhere unfossiliferous.

King Island

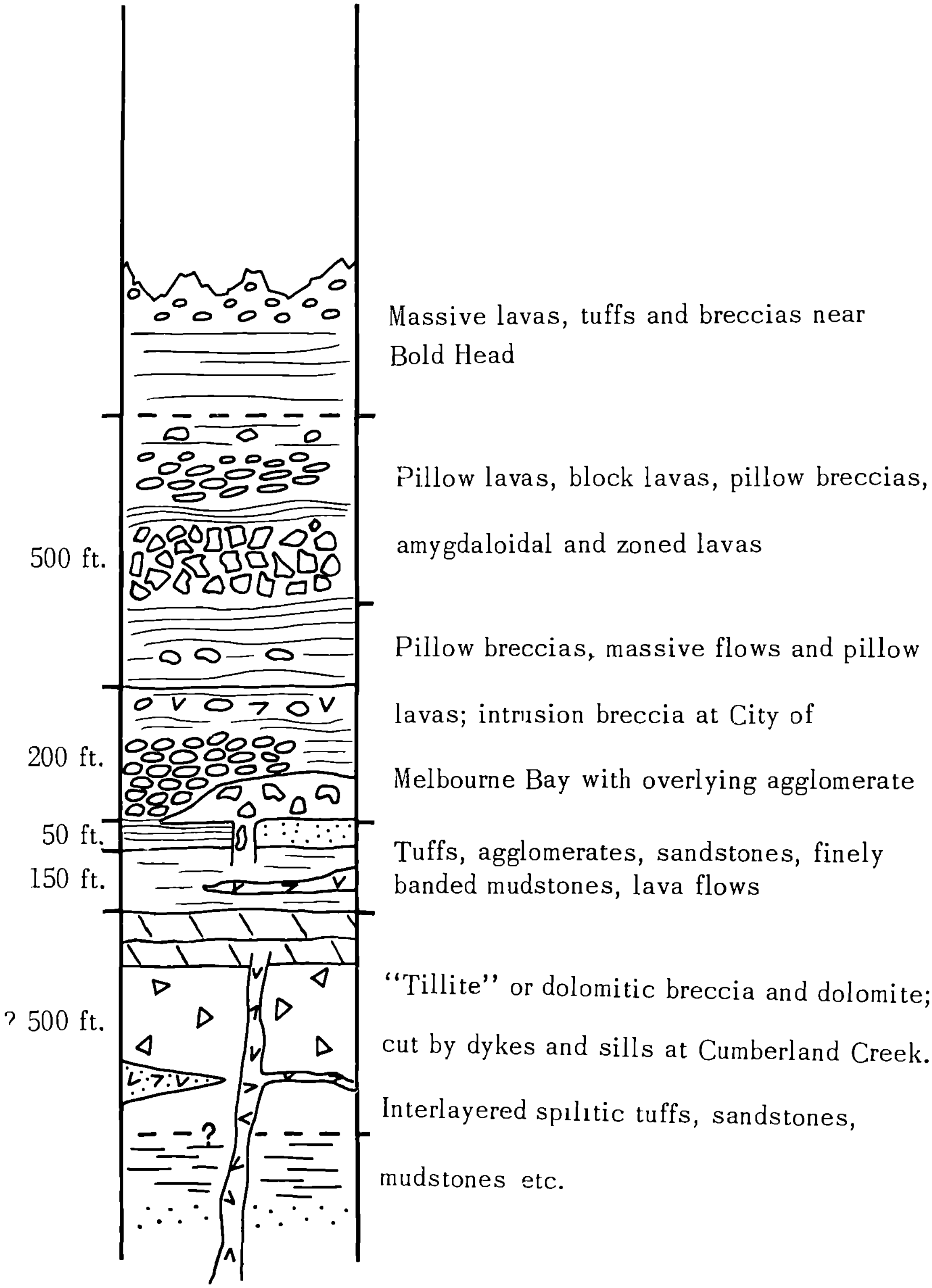
On King Island (Fig. 4) up to a thousand metres of volcanics conformably overlie dolomite breccias (the tillite of Carey, 1947b) and finely banded mudstone (varves ?). Mudstones are interbedded with lavas near the base of the volcanic succession (see Fig. 5). Individual lavas vary in thickness from over 30 m. to about 30 cm.,

Fig. 4: Geological sketch-map of the south-east of King Island, showing the distribution of spilitic volcanics.



144°05 E
NARACOOPA
FRASER BLUFF
N
Gabbro
BARRIER CREEK
CUMBERLAND CREEK
CAMBRIAN? Volcanics
Tillite?
PRECAMBRIAN? Sandstones, banded mudstones, tuffs
Granite
Gabbro
CONGLOMERATE CREEK
Shower droplet rock
50
50
40°00 N
YARRA CREEK
CITY OF MELBOURNE BAY
60
0 4000 8000 FT
0 120 240 M
50
GRASSY
GRASSY RIVER
BOLD HEAD
KING ISLAND SCHEELITE OPEN CUT

Fig. 5: Generalised geological succession in south-eastern King Island.


Massive lavas, tuffs and breccias near Bold Head
Pillow lavas, block lavas, pillow breccias, amygdaloidal and zoned lavas
500 ft.
Pillow breccias, massive flows and pillow lavas; intrusion breccia at City of Melbourne Bay with overlying agglomerate
200 ft.
50 ft.
150 ft.
Tuffs, agglomerates, sandstones, finely banded mudstones, lava flows
"Tillite" or dolomitic breccia and dolomite; cut by dykes and sills at Cumberland Creek.
? 500 ft.
Interlayered spilitic tuffs, sandstones, mudstones etc.
?

and vary considerably in their field characteristics. The chief types are as follows:

(a) Pillow lavas.

(b) Block or aa lavas (brecciated pillows ?).

(c) Isolated-pillow breccias, broken-pillow breccias, and broken-flow breccias.

(d) Zoned or pahoehoe lavas.

(e) Amygdaloidal lavas.

(f) Massive lavas.

(a) Pillow Lavas consist of thick masses of pillows in contact. The shapes vary between spheroidal, bun-like, and highly irregular, bulbous sacs, the spheroidal bodies being common only in the pillow breccias. Parts of the flows consist of bulbous sacs still connected by short necks; these sacs, and the isolated pillows, are flattened perpendicular to bedding, giving characteristic patterns when seen in plan and section (Plate 1, Nos. 1 and 2; Plate 2, Nos. 1 and 2. Individual bodies vary from about 6 cm. to 1½ m. in diameter, and a wide range in sizes may be visible in one outcrop (Plate 1, No. 2). Many individual pillows are deformed by contact with their neighbours but without cracking. This moderate deformation must have taken place while the pillows retained a fairly low viscosity and is presumably due to the weight of overburden in

Plate 1 No. 1:- Pillow lava, looking down dip, King Island, south of City of Melbourne Bay.

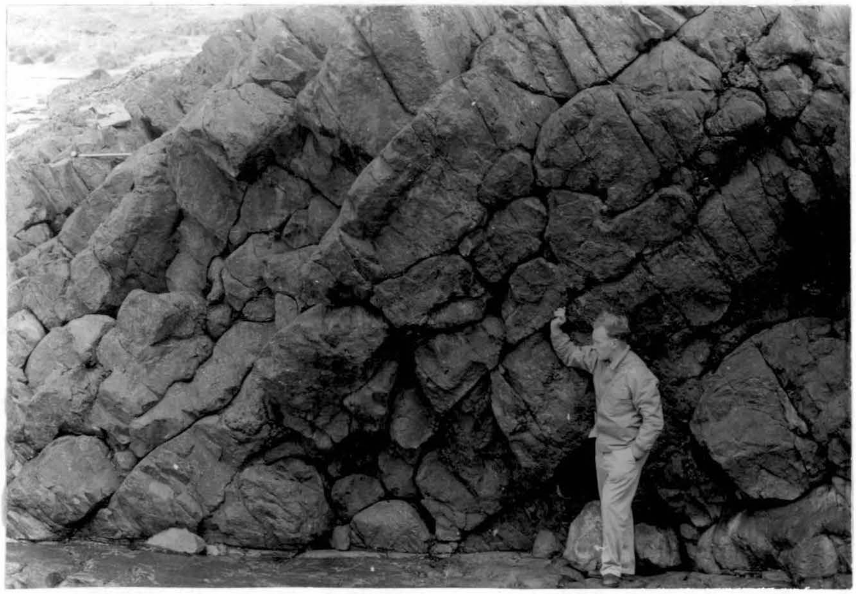


Plate 1 No. 2:- Pillow lava, looking along strike, same locality.

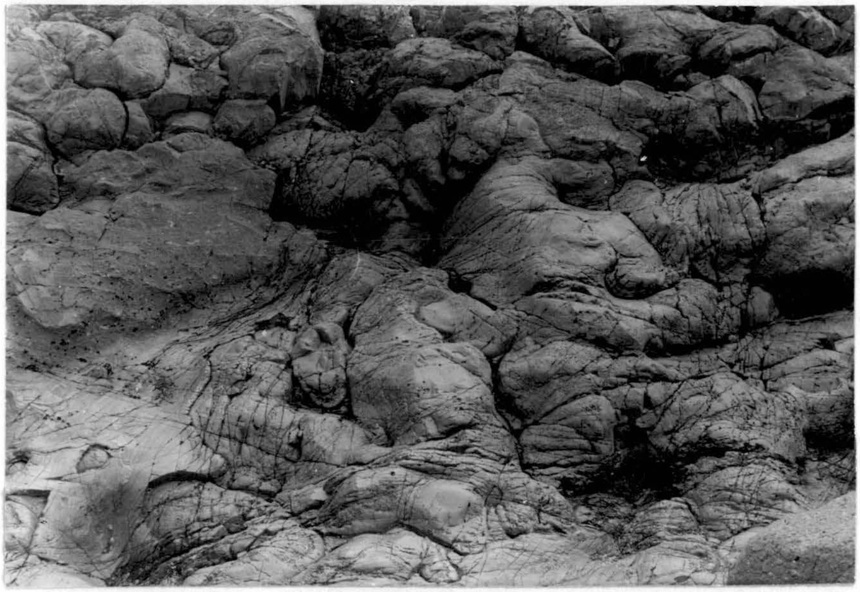


Plate 2 No. 1:- Pillow lava, looking down on bedding surface, same locality as Plate 1.

Plate 2 No. 2:- Large pillow sac, looking obliquely at bedding surface, same locality.

combination with continued movement of the lava flows. These pillow lavas are in no way exceptional when compared to classical text-book forms. As far as the internal structure of the pillows is concerned, there appear to be two main types in the pillow lavas, the second also occurring in the pillow breccias.

Type 1 Pillows

This type, shown in Plate 3 No. 1, is the most common and consists of an outer chilled margin generally 1-3 cm. thick, an inner zone of amygdaloidal spilite 3-9 cm. thick in which the amygdules are elongated radially, and a core forming most of the pillow (32407, 32401, 31867).* In the centre, ~~the~~ variolites up to 2 cm. diameter may be clustered.

Seen under the microscope, the chilled margin is dark-brown and weakly birefringent; it shows sheaf and radial structure and small circular patches (up to 0.1 mm. diameter) of lighter colour in which albite crystallites have formed. A few mm. in from the outer margin are found chlorite amygdules and very rarely albite phenocrysts up to 0.3 mm. long. The rims vary from 3-4 cm. to ½ cm. in thickness and the thickness appears to be independent of the size of

---

* Numbers refer to specimens held in the Geology Department, University of Tasmania, except where otherwise stated.

Plate 3 No. 1:- Typical pillow type I, from near Conglomerate Creek, King Island. The hand lens is 3.5 cm. long.

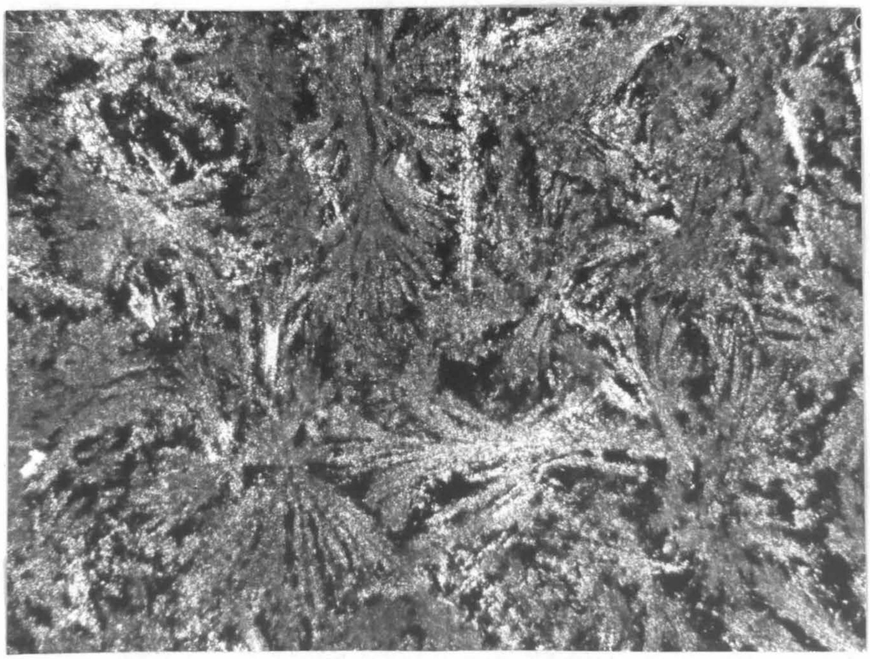


Plate 3 No. 2:- Groundmass of the amygdule zone in a type I pillow (31867b). X105.

the pillow.

The amygdaloidal zone (2-4 cm. thick) consists of similar dark brown glassy material but with larger patches containing albite and a little chlorite. The bundles and sheaves of incipient crystallites (?) are very pronounced (Plate 3 No. 2). The radial amygdules, up to 2 cm. long, contain chlorite and/or calcite with thin rims of very fine-grained epidote granules and/or dark ferruginous material. This material imparts a red colour to some of the amygdules. Pyrite cubes and splashes are common, particularly near these red amygdules.

Inside the amygdaloidal zone, the bulk of the rock still consists of almost crystalline bunches and sheaves. Rare amygdules consist of chlorite and/or epidote and clinozoisite. Occupying several per cent of the volume are groups of thin albite laths and some augite crystals (and occasional altered olivine crystals). The groundmass becomes more and more crystalline approaching the core which has typically basaltic texture, consisting largely of needle-like albite laths (up to 0.5 mm. long) with augite partly converted to chlorite, and interstitial magnetite and chlorite. A few phenocrystic albite laths (3 x ½ mm.) are found and several of these are distinctly curved (through 10^{o} to 15^{o}). Donnelly (1963) has described curved (up to 25^{o}) albites in quartz keratophyres from the Virgin Islands and he suggests this curvature indicates that the crystals were albite before the flows had solidified.

Very little matrix occurs between the pillows. It consists mainly of chlorite aggregate with a few thin albite laths, some of which are replaced by a very fine-grained brownish mineral (sericite, clay ?). The rock is veined and partly replaced by granular epidote.

Some pillowy flows have a skin 5-10 cm. thick which consists of greenish grey, finely banded, glassy tuff (see bottom left-hand corner of Plate 2, No. 1). This material (32426, 32405, 32404) appears to be composed of lenses and streaks of glass with blebs of devitrified (feldspathic ?) glass, crystals of epidote and calcite and quartz crystal fragments. Its origin is uncertain, but it may have formed by the flaking away of the thin glassy rims that probably formed around the pillowy flows; this material, distributed and broken off by current action and movement of the flows, probably settled back onto the flow surface as a detrital glass "sediment". In places this banded tuff completely encloses single pillows (e.g. Scott, 1951a, Plate 1, Figure 2). Scott (1951a) has given a chemical analysis of this material (Table 2 No. 6).

Chemistry of the Type 1 Pillows

A type 1 pillow was studied in some detail by chemical analysis (Table 2). Samples were collected from the outer rim, about 1 cm.

Table 2: Pillows, King Island

	1	2	3	4	5	6
SiO_2	50.36	48.62	50.04	51.42	44.54	51.14
TiO_2	0.81	0.70	0.72	0.77	0.29	0.49
Al_2O_3	13.43	14.07	13.22	12.17	9.07	9.01
Fe_2O_3	5.13	5.33	3.66	3.67	1.90	2.32
FeO	10.74	9.81	10.42	9.81	7.16	3.99
MgO	5.75	5.52	7.01	5.89	19.67	12.49
CaO	6.28	7.48	5.84	8.64	10.24	14.34
Na_2O	5.00	4.58	4.87	4.32	0.20	1.67
K_2O	0.05	0.05	0.05	0.13	Nil	0.39
H_2O+	2.29	2.64	2.94	1.96	5.57	2.64
H_2O-	0.27	0.20	0.20	0.24	0.47	0.20
MnO	0.21	0.20	0.20	0.25	0.16	0.16
P_2O_5	0.15	0.16	0.23	0.18	0.16	n.dt.
CO_2		0.22	0.21	Tr.	Tr.	0.13
FeS_2		0.39	0.17	0.21	0.13	
Total	100.47	99.97	99.78	99.66	99.56	99.74

Table 2: Pillows, King Island (continued)

		1	2	3	4	5	6
qu							0.46
or		0.29	0.29	0.29	0.77		2.30
ab		42.31	38.75	41.21	36.55	1.69	14.13
an		14.05	17.69	14.06	13.43	23.85	15.94
di		13.34	14.02	9.95	23.35	20.60	42.99
hy	en	4.14	3.15	5.67	6.21	21.09	13.58
	fer	4.21	2.94	4.97	6.00	4.91	2.14
ol	forst	4.83	4.91	6.43	1.81	13.99	
	fay	5.41	5.05	6.21	1.93	3.59	
mag		7.44	7.73	5.31	5.32	2.76	3.36
calc			0.50	0.48			0.30
ap		0.35	0.37	0.54	0.42	0.37	
ilmen		1.54	1.33	1.37	1.46	0.55	0.93
Total		97.91	96.73	96.49	97.25	93.40	96.13

1. Outer rim, 0.75 cm., of pillow type 1, 300 m. south of City of Melbourne Bay, King Island (specimen 31867a).

2. Amygdule zone, 2 cm. thick, same pillow (specimen 31867b).

3. Core, within 5 cm. of amygdule zone, same pillow (specimen 31867c).

4. Centre of core, same pillow (32401).

5. Outer 10 cm. of pillow type 2, 350 m. south of City of Melbourne Bay, King Island (specimen 32402).

6. Tuff between pillows (type 2), from Scott (1951, p. 122).

Analyses 1-5 by Dept. of Mines, Tasm.

thick, and from the amygdule zone, about 2 cm. thick; the core which forms the bulk of the pillow, was sampled at the centre and adjacent to the amygdule zone (Plate 4, No. 1). No analyses were made of the 1 mm. thick dark-green (chloritic ?) skin on the pillow. The striking feature of these analyses is the lack of variation across the pillow. Silica and lime are somewhat higher, and alumina, total iron, soda and water rather lower, in the core compared to the marginal zones but the variations aresmall.

Zoned pillows from other parts of the world show very marked variations though the chemistry of these is by no means uniform. Slavik (1928, from Vallance, 1960, p. 35) and Vuagnat (1946, 1949) have described margins rich in iron, magnesia and water and low in silica, lime and alkalies compared to the core and Nicholls (1959) has shown that in a zoned "sac" from Wales the margin is relatively rich in lime, magnesia, alumina, titania, total iron and water and poorer in silica, soda and carbon dioxide. Similarly, Vallance (1960, p. 36) found marginal zones rich in total iron, lime, water and titania and low in silica and soda in a pillow from New South Wales. Hopgood (1962) demonstrated in one pillow a concentration of silica between the core and outer margin and showed depletion of soda in the outer 1-2 cm. zone and local enrichment of soda between 2 and 4 cm. from the outer rim.

Plate 4 No. 1:- Close up of pillow type I of Plate 2 No. 2, showing positions of samples.

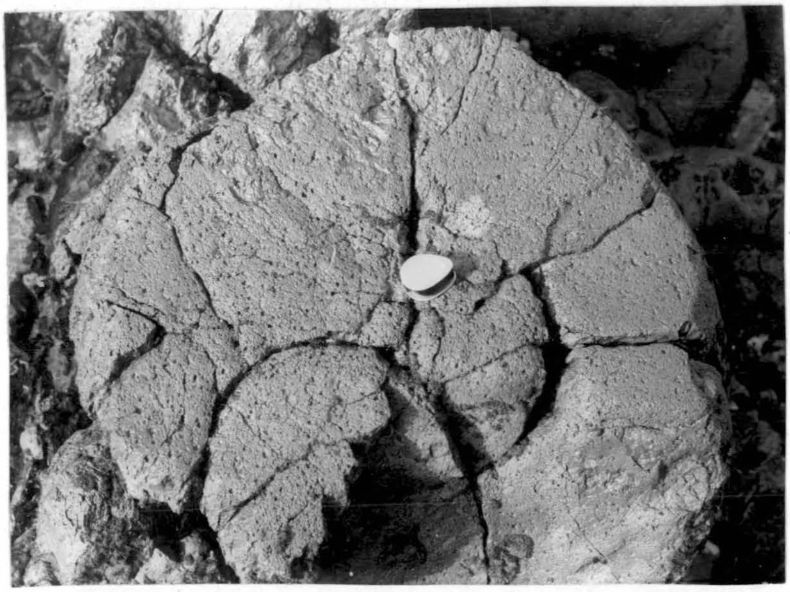


Plate 4 No. 2:- Typical pillow type 2, from near Conglomerate Creek, King Island. The hand lens is 3.5 cm. long.

The only reliable example known to the writer in which only very small variations occur, such as evidenced by the King Island rocks, is from the Welsh volcanics described by Nicholls (1959). This "homogeneous" sac contrasts sharply with his zoned sacs.

Bailey and McCallien (1960, p. 370) quoted two old analyses from Teall and Ransome, for pillows from Scotland and California respectively. Both are from the outer margins of the pillows and both are high in soda.

Most authors believe the chemical variations developed during cooling of the pillow. Hopgood and Bailey and McCallien believed sodium diffused outward toward the chilled margin, crystallising in albite inside the chilled skin. Nicholls suggested that the outer zones were metasomatised differently from the core zones during the late crystallisation stages, the outer zones being attacked by Ca-Mg-Fe solutions and the inner by Na-Si solutions. These two types are supposed to represent immiscible phases of the silicite melt. The existence of chlorite and albite emulsions was also suggested by Vuagnat (1946).

Some variations in a pillow are to be expected. The pillow probably develops a chilled margin almost immediately after extrusion and this probably acts as a semi-permeable membrane. The temperature gradient established across the pillow will cause water and volatiles to diffuse toward the margin. When volatiles separate as bubbles they will also travel outwards, becoming trapped near the rim, possibly as radially arranged amygdules. The core will thus be depleted of water, soda and potash. This theoretical picture admirably explains the distribution of soda in the Franciscan pillow and also that in the King Island pillow. The common relative enrichment in Ca, Mg and Fe in the rims may be due to reactions between the enclosing fluids and the pillow. The margin is probably not impermeable and may lose soda and other material by leaching and/or may become enriched in Ca, Mg and Fe derived from circulating gases. Chloritisation and calcium enrichment are particularly noticeable in the tuffs of King Island and Vuagnat (1953) has described intense chloritisation of pillow margins and tuffs from Mont Genevre, Switzerland. The source of the metasomatising solutions on King Island is discussed later.

In the analysed King Island pillow metasomatism from external solutions appears to have been of little importance and this may be due in part to the fact that it is embedded in a thick mass of pillows which may have prevented much contact with surrounding fluids, either sea water or volcanic gases.

Type 2 Pillows

The second type of pillow forms isolated, rare, pillowy flows, generally associated with pillow breccias. The pillows are commonly spheroidal and have a crudely defined internal zoning. A rather vaguely defined chilled margin envelops a featureless or a highly variolitic zone (variolites up to 2 cm. diameter); this encloses more featureless material that contains a core zone, 5-10 cm. across, with concentric bands of pale mineral. If the variolite zone is missing, the pillow is largely featureless.

The chilled margin (32402) and the featureless zones are composed of a brownish, very fine-grained felted mass of somewhat fibrous material (tremolite ?) containing olivine-like crystals pseudomorphed by chlorite. Tiny tremolite needles project into these crystals from the margins. The variolites (31828), where present, show chloritised olivine crystals in crudely radial sheaves of fine-grained, brownish material with crystallites that appear to be largely pyroxene. The core zone consists of similar, fine-grained tremolite and chlorite with concentric bands a few mm. thick of soft, almost white material that appears in thin section as a brownish very fine-grained aggregate of fibrous materials. X-ray powder photographs indicate a mixture of tremolite and possibly chlorite or talc. Very similar flows without any zonal arrangement are common (Plate 4, No. 2).

An analysis was made of part of the outer 10 cm. of a pillow of this type (Table 2, No. 5) and may be compared to the picritic spilite of Table 5 (No. 1). The margin has apparently undergone some enrichment in H_2O and MgO, probably present in the tremolitic material. This alteration is analogous with the chloritization of the enclosing tuffs. Pillows of the second type occur in pillow breccias and transitional rocks with partly broken, highly deformed pillows (Plate 5, No. 1). They do not form thick pillowy flows like type 1.

Origin of Pillow Lavas

The origin of pillow lavas is a subject that has seldom been discussed in detail. Daly summed up the position in 1933 when he stated (p. 419) that "the structure has been connected with the development of the 'spheroidal state' at the contact of water and hot basic lava, but no one has made the matter clear". Wilson (1960) presented a summary of previous views on pillow formation after examining pillows in Keewatin spilites and recently Snyder and Fraser (1963, a and b) have given an even more detailed summary of previous work. Most discussion has centred around the problem of whether pillows necessarily develop in water and Wilson concludes that water is necessary (a) to provide rapid cooling, (b) to develop the stratified rocks that are "commonly interbedded with pillows" (p. 100), and

(c) to allow a crust to develop on one layer of rounded pillows before the next develops. The only valid point of these three is probably (b), though Wilson is probably correct in arguing that water forms an essential part of the process. Snyder and Fraser (1963, p. C2) believed "that the general absence of pillow structure in most subaerial flows, however, militates against the general effectiveness of all subaerial mechanisms. We believe that the juxtaposition of hot fluid magma and a cold fluid, generally water or muds, is generally necessary for pillow formation".

Many textbooks (e.g. Holmes, 1944, p. 449; Rittmann, 1962, p. 70; Tyrell, 1931, p. 51) describe pillow lavas as the subaqueous equivalent of pahoehoe flows, probably under the influence of descriptions by Anderson (1910, p. 633) of pahoehoe lava pouring into the sea on Savii, in the Samoan Islands. Snyder and Fraser (p. C3) have pointed out that the globular bodies (not real pillows) seen by Anderson were merely budding pahoehoe flows and formed both above and below the waterline. They also point out that there is no direct relationship between pahoehoe and pillow lavas. Because it has not been proved that water is a necessary part of pillow formation, there have been several recent papers suggesting subaerial processes. Osborn (1949) produced "cellular" (i.e. globular) bodies on a small scale by pouring molten fluosilicate glass into a tank, and he suggested the globules were convection cells. He indicated that pillow lavas might form in the same way and cited descriptions by Green (1887) of spheroidal bodies that formed subaerially in flows on Mauna Loa. These Hawaiian spheroids apparently only developed as the lava fell from a low wall into the sea and under this condition of free fall, spheroidal forms are very likely to develop. Such circumstances, however, must surely be exceptional. Osborn's experimental globes seem to be closely related, as he himself noted, to polygonal cracking and probably originate by incipient columnar jointing, and Snyder and Fraser (1963b, p. C3) suggest that very few pillows can be found in the Hawaiian flows.

Yagi (1960) believed the pillow flows of the Nemiro Peninsula, Hokkaido, Japan, formed by a mechanism similar to that envisaged by Osborn.

However, the weight of geological opinion falls heavily on the side of an aqueous origin and this is supported by theoretical considerations. It is suggested here that the process by which masses of pillows form is the result of interaction between the opposing forces of surface tension and gravity, and that a fluid medium (mainly water) is a necessary component. In magmas of low viscosity in contact with air the surface tension is insufficient to form the minimum surface energy form (the sphere) because of gravitational effects. The density contrast between molten basalt and air is about 2,800:1, but in water the density contrast is reduced to about 2.8:1, and in the case of salt water with ash or mud in suspension, the contrast is further reduced. The fact that many of the best examples of pillows appear to have formed in, or rolled into, muds is probably related to the small density contrast between mud and magma. Under low density contrast conditions it is possible for a spherical form to develop and for it to continue growing as long as magma is being supplied. The pressure of lava required to enlarge the sphere a given volume (i.e. the pressure across the radius, P) will decrease as the radius increases, such that $p=\frac{2T}{r}$, where T = surface tension and r = radius of sphere (Kingery, 1960, p. 195). At some stage the pillow would separate from its feeder source and possibly roll away or alternatively would "bud" or "neck" into two or more pillows. Other factors such as movement of the source, slope of the floor, and of course, the pressure of the lava, will interact with gravity to control the size and shape of the pillows.

That surface tension is a vital factor has been implied or stated by Fuller (1940) and Lewis (1914) in their descriptions of "gigantic" or "giant" emulsions, and also by Wilson (1960) who believed that lava globulated "in the same way as oil globulates when poured into cold water" (p. 101). Snyder and Fraser (1963a, p. B19) supported the idea of an emulsion but believed that surface tension could scarcely contain large globules of lava. However, as just pointed out, it could do so if the density contrast between pillow and medium was very low. The globules, of course, would only form in the first place because of surface tension effects.

Wilson (1960) suggested that the development or non-development of pillows is controlled by the size of the flow, the temperature of the lava, and the velocity of the lava prior to consolidation. From observations at King Island, the size of the flow appears to have little effect as some pillowy flows are only as thick as the width of the pillows and yet others contain many pillows. The temperature would affect the viscosity and the surface tension and thus affect the ability to globulate. The surface tension and density also vary with composition and empirically there is clearly a compositional control, for pillows are relatively common in basic lavas but rare in intermediate and acid types (see Snyder and Fraser, 1963b, p. C5).

The control by the rate of flow is less easy to evaluate but may well be one of the major factors, not only in controlling the size of pillows as already described but in controlling the behaviour on breaking through rock into water.

Rittman (1962, p. 71) suggests that the size of pillows is controlled by the opposing factors of lava pressure and rate of cooling, and the chilling effect of water is quoted in many text-books as the main cause, or one of the causes, of pillow lava. The lava pressure is clearly important, but the rate of cooling probably relatively unimportant. The rate could control the thickness and hence the strength of the rim, which in turn might affect the growth of the pillow, but the cooling rate for a particular field of pillows is likely to be much the same and could hardly account for the wide range in sizes observed, for example, in the King Island pillows (Plate 1, No. 1). The rate of cooling might control the maximum size of pillow that could form in a given set of conditions. Because the ability of the cooling sac to withstand fracture under conditions of quenching is inversely proportional to its size (Kingery, 1960, p. 636), there is likely to be a size above which disruption of the pillow would be inevitable.

The formation of pillows is apparently independent of the depth of water, as pillows have been observed forming in shallow water and recent pillow fields have been photographed in water up to 14,400 ft. deep (Moore and Reed, 1963). The hydrostatic pressure on ocean floors is likely to influence the form and size of the pillow, but not prevent their formation. Most pillows probably develop from underwater eruptions and it should be possible to distinguish these from pillows developed when subaerial lava passes into bodies of water, by the contiguity of subaerial and subaqueous structures in the flows.

To summarise, the development of pillow lavas is probably as follows: Lava of low viscosity welling from the vents forms pillows immediately on contact with water or saturated mud, giving rise to pillow flows. If the rate of flow of the lava is initially high, a large fluid sac with a thin chilled skin forms, out of which burst tongues of lava to form pillows. The sac itself tends to form bulbous tongues and lobes and may break up into crude pillows or pillowy masses connected by "necks". Once a thin, chilled skin has developed the rate of cooling falls rapidly and the sacs stay plastic for a considerable period, during which they may be rolled by movement of lava, or flattened.

Ascrystallisation commences, volatiles are released from solution. In some pillows (e.g. Type 1 from King Island) these partly escape through the semi-permeable skin and are partly trapped in the pillow to form amygdules; however, in other pillows the volatiles are unable to escape and the build-up of vapour pressure may cause disruption of the pillows to form block lavas and broken-pillow breccias.

(b) Block Lavas ('aa' flows of Scott, 1951d) are relatively rare and occur within areas of pillow lavas, into which they grade. These lava breccias (Plate 5, No. 1 and Plate 6, No. 2) consist of angular to subangular blocks of lava up to 30 cm. across in a very sparse

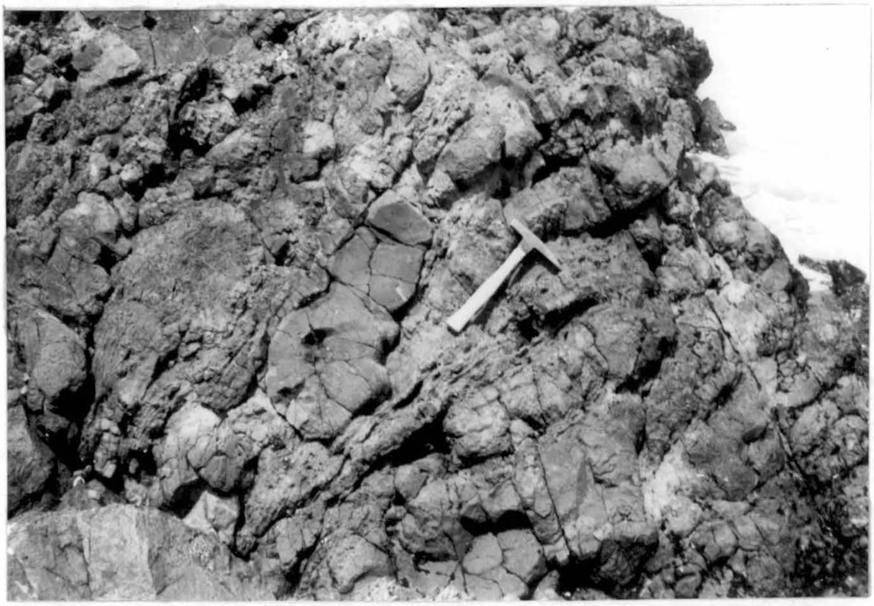


Plate 5 No. 1:- Deformed pillows, looking north along strike, type 2, 500 m. north of City of Melbourne Bay.

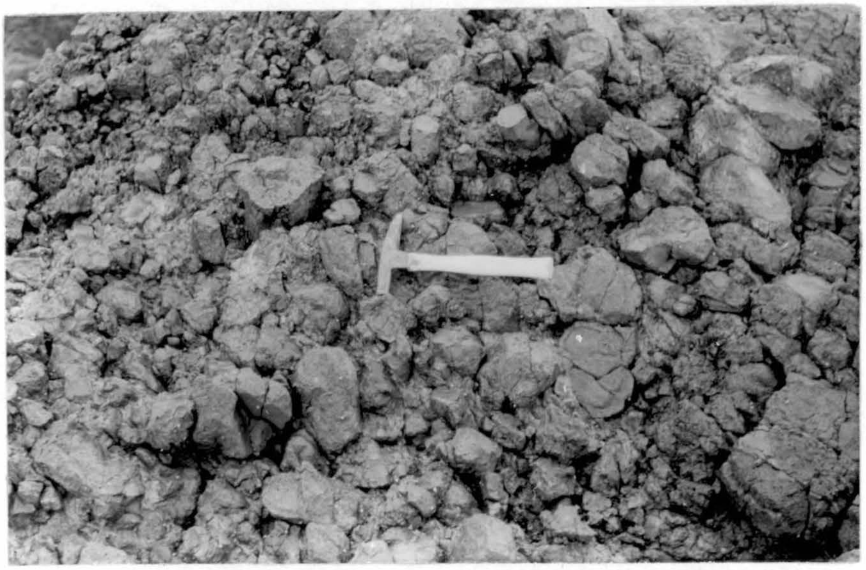


Plate 5 No. 2:- Blocky lava, 500 m. north of Conglomerate Creek, King Island.

Plate 6 No. 1:- Blocky lava, 500 m. north of Conglomerate Creek, King Island. Field is approximately 6 metres across.

Plate 6 No. 2:- Disrupted pillow type 2, 500 m. north of Conglomerate Creek, King Island. Silver coin is 2.5 cm. radius.

spilitic matrix. They appear to be derived from the cracking and disintegration of pillows; for various stages of the break-up of pillows may be seen. Occasional cracked and deformed pillows occur within the breccias, a feature noticed also by Scott, who figured a budded lava within block lava, the budded sections being strongly cracked (her Plate 2, Figure 1).

In Plate 5, No. 2, and Plate 6, No. 2, a rim of finer grain can be seen around several of the blocks, indicating that disruption took place at an early stage and involved slight chilling of still plastic material. In Plate 6, No. 2, a group of chilled blocks are clustered together to form a crudely defined spheroidal body that may have been a pillow. It is possible that the pillow formed a chilled skin which, because of increasing vapour pressure within, failed in a brittle manner and allowed access of solutions (or water) into the still molten core. This then formed a series of second-order pillows with chilled margins. In many cases, disturbance of the cluster of new pillows may have resulted in typical blocky lava.

Some pillows may have broken up by explosion due to internal vapour pressure. Washington (1909), who saw the submarine eruption of 1891 near Pantellaria, described subspherical bombs, red-hot, that were ejected above the surface of the water.

These were up to 3 ft. in diameter and "many ran hissing over the water". After a while, they exploded and the fragments sank. These explosive bombs, or pillows, were derived from rapid ejection, were of large size, and had a high gas content. The bombs probably exploded during the build-up of high internal vapour pressure against the solid skin. The pillow cores would probably fracture rather than spatter or flow provided the relaxation time of the spilitic melt was longer than the period of the explosion. As the latter is so short, fractures might form at quite a high temperature. Discussions by Carlisle (1963) and Barth (1962, p. 137) indicate that natural basaltic lavas vary in viscosity from 10^5 to 10^2 poises. For the spilites, the figure of 10^2 seems probable but in the pillows would have risen to, say, 10^5 poises. For such material a relaxation time of , say, 10^{-3} to 10^{-5} seconds might be reasonable. An explosion would probably take about 10^{-3} seconds to pass through a pillow and hence the material could fracture. Basalts have a viscosity of about 10^5 poises at temperatures between 825°C and 950°C, but for "wet" basalts the temperature might well be lower, say 750° C - 800°C.

It is easy to imagine that exploding pillows, either near the surface or at considerable depths, could produce local fields of blocky lava containing some "unexploded bombs".

Another possible cause of block lavas is cracking as a result of cooling. Shrinkage stresses set up during cooling may be sufficient to induce fracture, and radial cracks of this type have been figured by several authors (e.g. Rittmann, 1962, p. 37; Yagi, 1960, p. 914). The resistance of the pillow to fracture (its "thermal endurance") is dependent on several factors (Kingery, 1960, p. 636) but it is clear that the thermal endurance is inversely proportional to the size of the body. Hence larger pillows are more likely to crack than smaller ones in a given set of conditions. So few pillows on King Island are cracked that this conclusion could not be tested empirically and the writer is unaware of any other study that might confirm it. If cracked pillows were mobilised or disturbed after consolidation, perhaps by movement of adjacent molten lava, then block breccias would result. However, little evidence of this process is visible in the blocky lavas of King Island, and the pillows, even the largest, possessed remarkably few cracks. Probably the slow rate of cooling in the cores of the unexploded pillows in a heated water environment, allowed viscous relief of the cooling stresses.

The petrography of these lavas has been described by Scott (1951, a). They consist entirely of a dusty, very fine-grained aggregate of tremolite (?) needles with phenocrysts up to 1 mm. long of chlorite pseudomorphing olivine.

They are picritic basalts identical with the pillows of type 2 and also to the amygdaloidal and zoned flows but are distinctly different from the type 1 pillow lavas in mineralogy and composition.

(c) Isolated-pillow, broken-pillow, and broken-flow breccias.
The first two names are taken from Carlisle (1963) who has described pillow breccias in Triassic andesite-basalts on Quadra Island, British Columbia. The first type consists of whole, but somewhat deformed, pillows separated by tuff and the second consists of disaggregated pillows in tuff. There are gradations between these types and between isolated-pillow breccias and pillow lavas.
The broken-pillow breccias merge to broken-flow breccias in which fragments of thin flows are enclosed in a tuff matrix. These breccias are particularly well exposed for a mile or so south of the City of Melbourne Bay; they overlie pillow lavas near the base of the sequence and form units several tens of metres thick.

Typical breccias are shown in Plate 7, Nos. 1 and 2. These consist of "beds" of breccia interleaved with continuous flows (Plate 7, No. 1). The blocks vary greatly in size but are mainly about 15 cm. across, are subangular and chaotically distributed in a tuff matrix. Scattered through the breccia are whole pillows of type 2 (Plate 8, No. 1) and some large angular fragments.

Plate 7 No. 1:- Spilitic breccias interbedded with thin flows, 300 m. south of City of Melbourne Bay, King Island.

Plate 7 No. 2:- Spilitic breccia with whole and fragmentary pillows, 250 m. south of City of Melbourne Bay.

Plate 8 No. 1:- A pillow type 2 in breccia, same locality as Plate 7 No. 2.

Plate 8 No. 2:- Isolated pillow breccia, south of City of Melbourne Bay.

Essentially these breccias appear to be broken-pillow types, containing only rare whole pillows. A type closer to isolated-pillow breccia is shown in Plate 8, No. 2, in which rather deformed pillows and some pillow fragments are scattered through a tuff matrix.

Breccia transitional to broken-flow breccia is shown in Plate 9, No. 1. In this, rafts of thin flows and fragments of flows are strung out ina very plentiful tuff matrix. In other cases, the flows are largely undisturbed (Plate 9, No. 2) while in others they are so torn up that they resemble broken-pillow breccias. The field evidence suggests that these thin flows have been intruded into the tuff, which was then saturated with water, steam, etc., and in many cases was unable to support the flows. These broke up, lensed out or balled up, to some extent settling into the soggy tuff. Breaking up of some small flows may have been assisted by the development of cracks due to cooling. These breccias merge into similar breccias resulting from break-up of pillows rolled into the tuff, fragmentation of the pillows probably occurring in the same way as for block lavas. A transitional type with a whole pillow and large irregular fragments of flows is shown in Plate 10, No. 1.

In thin section, the pillows, pillow fragments and flow fragments (32408, 32419, 32410) are similar to the block lava and pillow

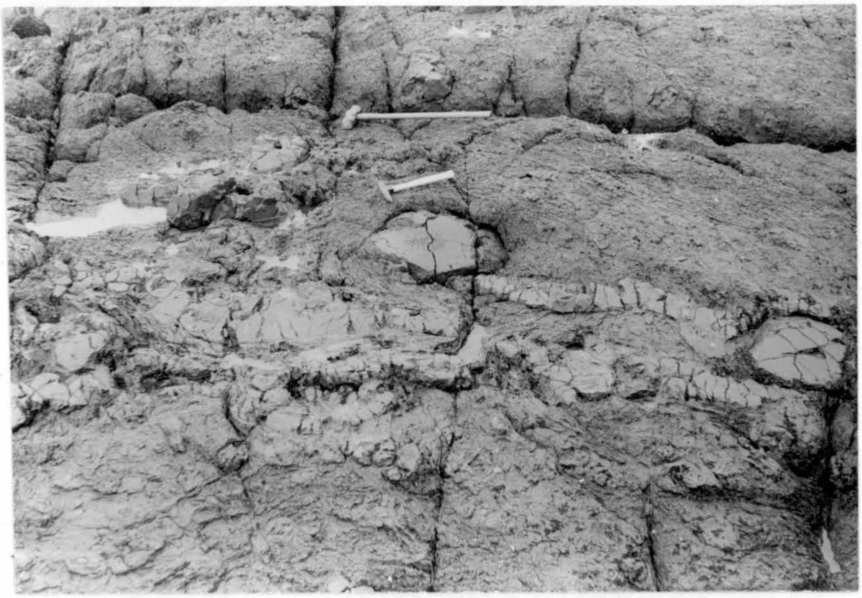


Plate 9 No. 1:- Broken-flow breccias, looking down dip, south of Conglomerate Creek.

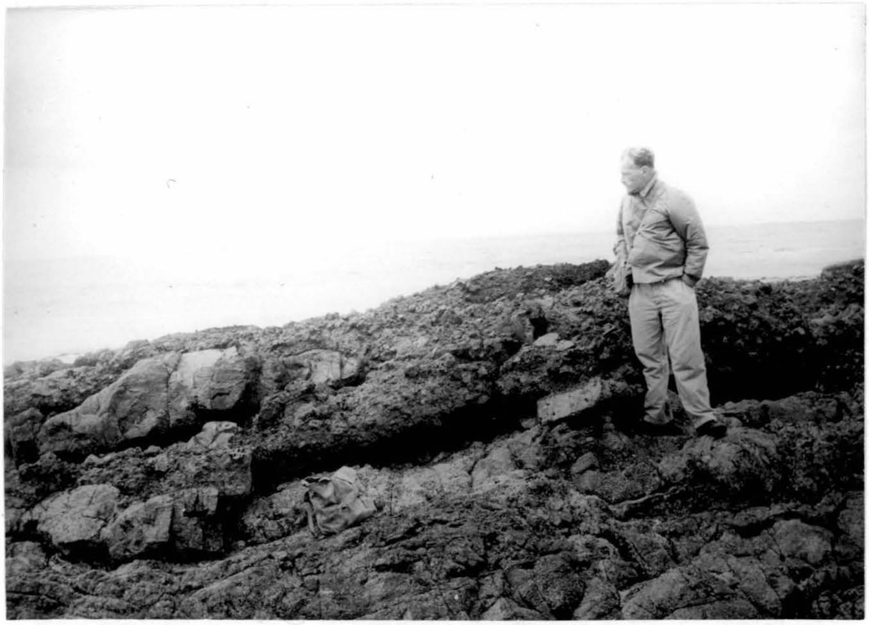


Plate 9 No. 2:- Thin flow in breccia overlying fairly massive flow, south of Conglomerate Creek.

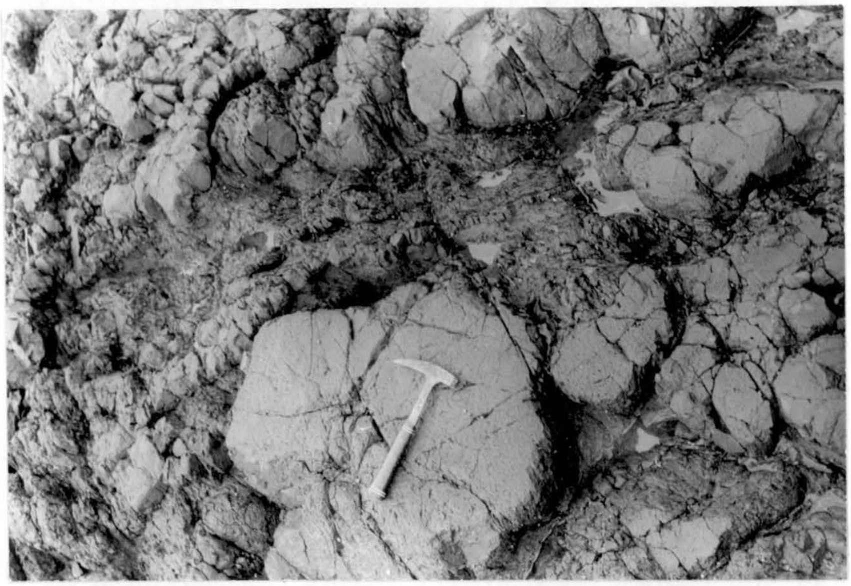


Plate 10 No. 1:- Breccia of fragmented flows and whole pillows, south of Conglomerate Creek.

Plate 10 No. 2:- Tuff matrix to spilitic breccias, 50 m. south of Conglomerate Creek. The hand lens is 3.5 cm. long.

type 2 previously described. Many slabs and fragments of pillows show on weathered faces a chilled rim (e.g. 32410) which in thin section is revealed merely as a darker coloured, more finely grained groundmass. This rim indicates the fragmentation was penecontemporaneous with extrusion.

The Tuff Matrix

As might be expected, the tuff matrix is contorted (see Plate 8, No. 2) and crudely foliated, the foliation wrapping around the blocks and pillows. Care has to be taken to distinguish between this primary foliation and a later, weakly developed, tectonic cleavage. Closer examination of the tuff shows it to consist mainly of irregular fragments up to 1 cm. diameter of greenish-grey material, many of the fragments having a thin white or bluish-white rim (Plate 10, No. 2). Some fragments are dark green and consist mainly of chlorite.

Thin sections show some variation in the tuff. 31817 consists of irregular blebs and lenses up to several cm. diameter of devitrified glass (?) with thin rims (Plate 11), the larger blebs containing almost circular masses up to 5 mm. diameter with a well defined dark brown rim (Plate 12, Nos. 1 and 2). The blebs and lenses consist largely of fibrous, very pale green, non-pleochroic chlorite and

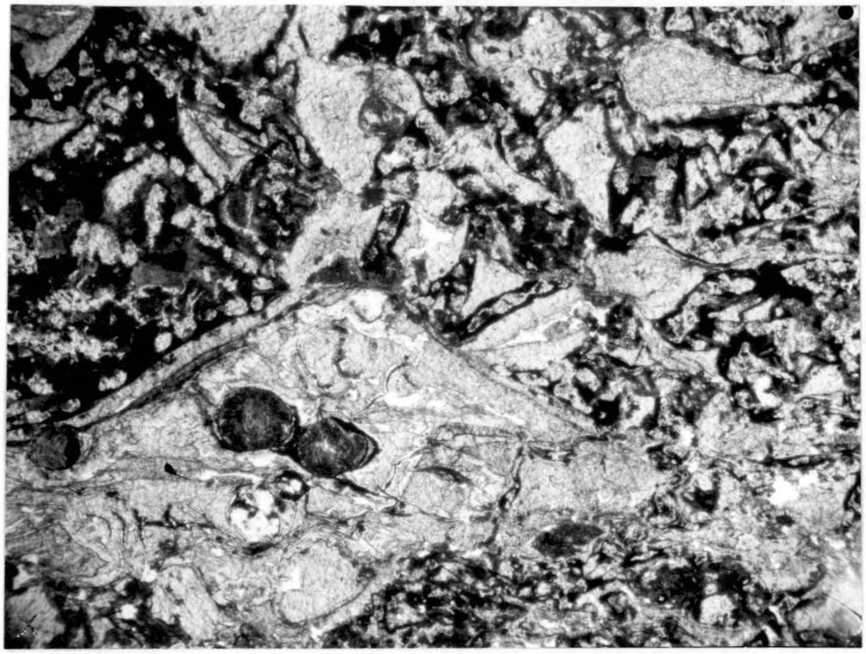


Plate 11 :- Portion of tuff in 31817. X20.

Plate 12 No. 1:- Irregular blebs of glassy material in 31817. X6.

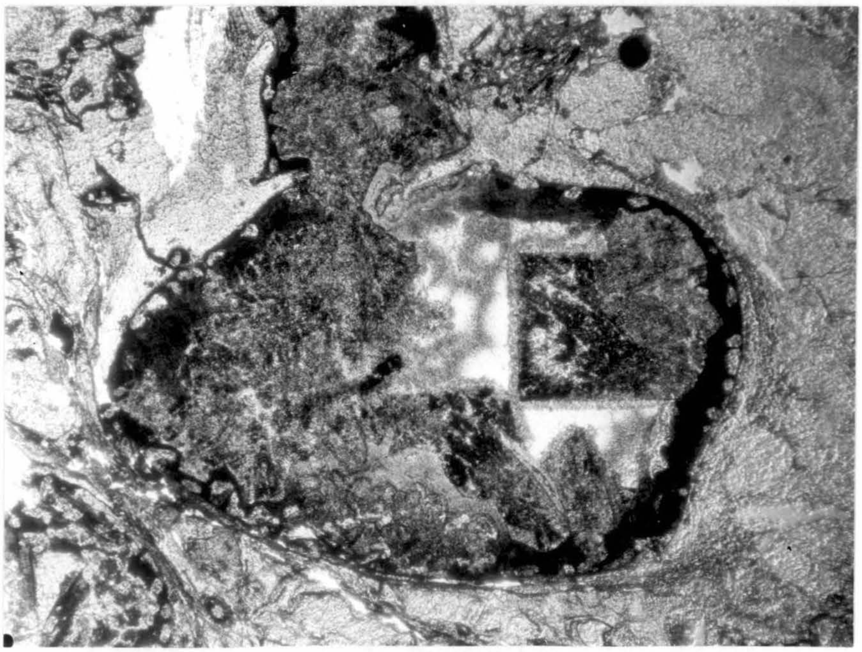


Plate 12 No. 2:- Enlargement of lower circular body of Plate 12 No. 1. X30.

some show circular, perlitic (?) cracks. The larger masses are very irregular in shape, with mainly convex margins, and appear to have been plastic at some stage. Smaller fragments forming a sort of matrix have in many cases concave margins and almost shard-like outlines (Plate 11). The circular masses, which may be amygdules or represent some immiscible phase, have dark brown, slightly birefringent, crudely fibrous rims with a sharp outer margin and an inner margin that is very irregular or crudely scallopped. The cores are apparently devitrified glass, in some cases containing patches of quartz, feldspar or chlorite. Other examples of tuffs with irregularly shaped particles are shown in Plate 13, Nos. 1 and 2.

In 31816, the tuff is largely composed of ovoid elongate globules from 0.5 to 0.3 mm. diameter. These have a thick, rather fuzzy rim of dark brown glass and an inner rim of very slightly birefringent pale brown glass within which are crystallites, some with high birefringence and high relief (epidote ?). Much of the brownish material mentioned is white in reflected light and is probably hydrogrossular. The core of these globules, often very insignificant, is of pale, almost colourless chlorite, in many cases mixed with granules of devitrified glass. One or two of the small globules are hook-shaped but the majority are elongated and streaked out in one direction; this imparts to the tuff the very crude foliation

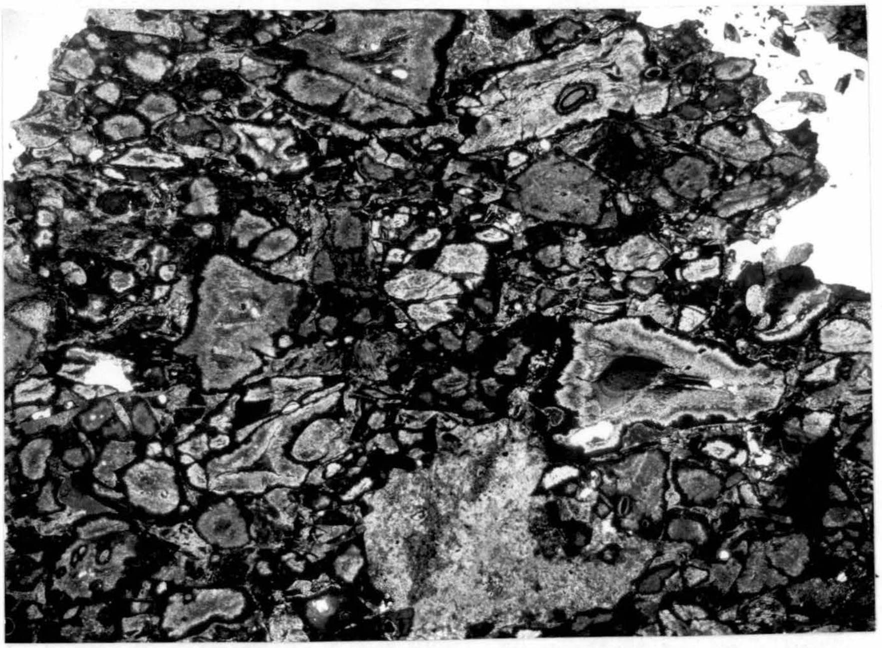


Plate 13 No. 1:- Tuff of specimen 32414. X10.

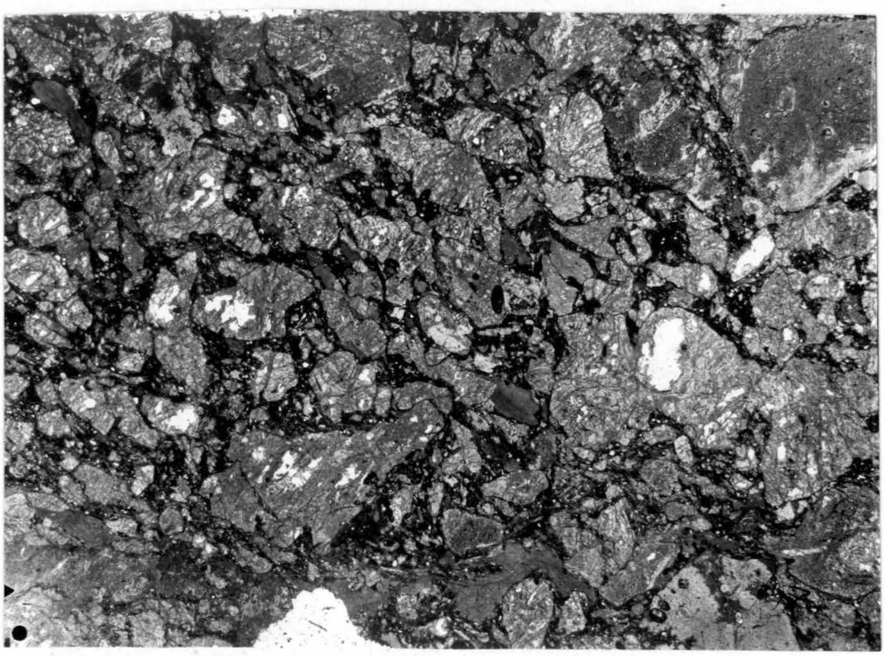


Plate 13 No. 2:- Tuff of specimen 32409. X5.

previously described. There are also elongate or irregular bodies of perfectly clear material that is almost isotropic but shows minute crystallites and patches of birefringent aggregate. These have the appearance of tachylite. Most of the glassy fragments appear to be slightly devitrified, partly chloritised basaltic glass. Some contain amygdule-like bodies rich in hydrogrossular and in some specimens (e.g. 5175) hydrogrossular has replaced most of the fragments and also occupies intersititial "pools".

The fragments generally are similar to the sideromelane described by Peacock (1926a, 1926b) from Iceland. Scott (1951a and Table 3) has analysed a sample of the tuff which the writer believes to be relatively low in chlorite and rich in hydrogrossular, and is similar to sideromelane (Peacock, 1926b, p. 57 and Peacock and Fuller, 1928, p. 371). It has an average refractive index of 1.69 according to Scott (1951a, p. 118). It is relatively higher in $Ca0$, Fe_20_3 and water, and low in $Mg0$ and $Fe0$ but is clearly not sufficiently hydrated to be a palagonite. Compared to Peacock's palagonite it is high in $Si0_2$, $Ca0$ and Al_20_3, and low in $Mn0$ and water content. The analysed tuff is rich in hydrogrossular compared to chlorite (note the relatively low $Mg0$ and high $Ca0$), is partly hydrated and most of the iron is oxidised. It might be described as a partly hydrated, hydrogrossular sideromelane vitrophyre. Analysis 2 in Tabe 3 is of the more

Table 3: Volcanic Glasses

	1	2	3	4
SiO_2	44.14	44.32	46.39	35.34
TiO_2	0.31	0.37	1.27	2.10
Al_2O_3	15.63	13.41	16.27	11.15
Fe_2O_3	5.45	2.67	1.35	10.28
FeO	0.93	7.83	9.96	2.19
MgO	2.75	15.98	9.77	6.52
CaO	20.44	6.90	13.00	7.01
Na_2O	0.80	1.28	1.40	0.16
K_2O	0.21	0.05	0.15	0.19
H_2O+	6.04	6.01	0.15	8.90
H_2O-	6.04	0.51	0.10	15.50
MnO	0.11	0.13	Tr.	0.22
P_2O_5	n. dt.	0.13	0.05	0.24
S				0.07
CO_2		Tr.		
Total	100.81	99.59	99.86	99.87
qu	4.38			
or	1.24	0.29		
ab	6.77	10.83		
an	38.44	30.70		
di	14.77	2.14		

woll		18.37	
hy	en		21.70
	fer		6.43
ol	forst		12.13
	fay		3.97
mag		2.46	3.87
hem		3.76	
ap			0.30
ilmen		0.59	0.70
Total		90.78	93.06

1. "Volcanic Glass" from King Island (Scott, 1951a, p. 122).
2. Vitrophyre from 200 metres south of Conglomerate Creek.

   Analyst: Department of Mines Tasmania.
3. Sideromelane from Gulfoss, Iceland (Peacock, 1926b, p. 57).
4. Palagonite rock from Hvalfjorour, Iceland (Peacock, 1926b, p. 66).

common chloritic vitrophyre. It is similar to sideromelane, but being chloritised, has high MgO and H_2O and low CaO.

Peacock (1926a) described "bleaching" (to palagonite ?) of the rims of sideromelane fragments and an analagous alteration might well be the development of white hydrogrossular rims on the King Island tuffs. The fibrous brown rim material just described has some similarities to Peacock's "fibro-palagonite" which he regarded as an incipient development of limonite and chlorite.

In 32421, the sideromelane particles have a thin rim (up to 0.05 mm.) of clear quartz (Plate 14) and the basaltic core is partly or wholly replaced by granular epidote. Quartz has replaced the brownish glass in some specimens. In others (e.g. 32414, 32417, 32406) many of the fragments are composed largely of tremolite needles and some contain chloritised olivine crystals. These types appear to represent a transition to fragmented pillows and flows.

The vitrophyres have many similarities to the "aquagene tuffs" of Carslisle (1963) and to the "palagonite tuffs" described from the subglacial volcanics of Iceland (Peacock, 1926a, b). They differ from the shard-tuffs that result from spalling of glass from the rims of pillows in that they contain globular particles; in any case, the King Island tuffs could hardly be derived from the pillow skins because of the abundance of tuff compared to the relatively rare pillows and

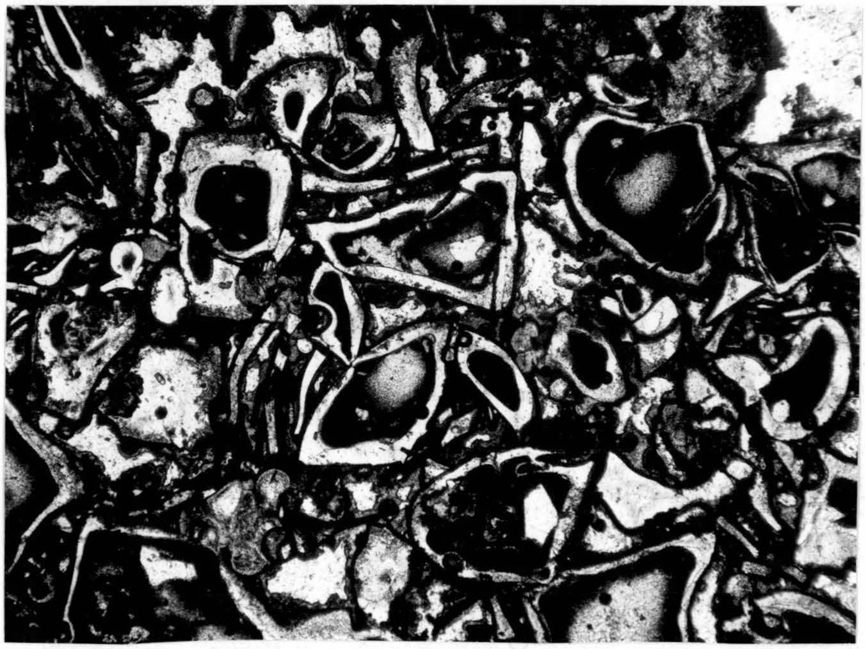


Plate 14 :- Quartz rims to sideromelane particles
in 32421. X40.

lava fragments. Carlisle believed his aquagene tuffs were formed entirely under water by expulsion from vents of basalt magma (not spilite) that immediately globulated, the globules either exploding, or sinking to the sea floor and then cracking and disintegrating. The resulting tuffs thus contain many shards with concave boundaries.

In the tuff fragments of King Island concave boundaries are relatively rare, many fragments being ovoid or elongate; they appear not to have fractured but to have undergone only viscous deformation, perhaps soon after settling on the sea floor. Carlisle made artificial concave shards by pouring molten basalt into cold water and his process seems entirely reasonable. For the King Island rocks, however, there are several factors which may have resulted in a rather different behaviour. First, the spilitic magma was probably relatively rich in water and at a somewhat lower temperature than a normal basalt; second, the high water content of the magma would have lowered the viscosity of the glass; and third, considerable quantities of steam and other gases must have been emitted, thus heating the water around the eruption centres and reducing the chilling effect. Heat transfer out of the tuff layers would be extremely slow by conduction and probably mainly took place by convection of the water both within the tuffs and above the volcanic rocks. Convection within the compacting tuffs, which contained lava flows and hot pillows and were penetrated by gases, was probably limited. Though initial quenching undoubtedly occurred, a phase of slow cooling seems likely and might well have been slow enough for the tensional stress to be relaxed by viscous strain.

Viscous deformation of the King Island tuff globules was probably partly due to compaction but may also have taken place during flowage of the tuff. The intrusion of thin flows, combined with surges in the pore water and local slumping, must surely have resulted in stretching and streaming of globules, particularly around larger globules and rock fragments. The foliation visible today is an expression of this globule elongation. Deformation of this phase probably continued until the globules had cooled below about $450^{\circ}C$, when the viscosity was probably about 5×10^{14} poises (Morey, 1954, p. 145-149; Bowen, 1934, p. 253) and the relaxation time about 10^{3} seconds.

An unusual rock, known as the "shower droplet" rock has been described by Scott (1951a). She believed this rock formed as "the result of the accumulation of small drops of lava or lapilli, which have dropped one on top of the other when almost, but not completely, solidified" (p. 117). Her photographs have been poorly printed and the outcrop and a close-up are figured in Plate 15, Nos. 1 and 2. The rock (32423) forms a bed nearly one m. thick among lava flows and itself appears to have a thin chilled margin (see Plate 15, No. 1). It consists of spherules, rods and irregular fragments of picritic spilite (Plate 15, No. 2), the irregular fragments being up to several cm. across, the rods several cm. long and generally curved, and the spherules, which form most of the rock up to 1 cm. diameter. Some spherules show a very thin chilled margin. This rock appears to be a cross between small scale pillow lava and the sideromelane tuffs and is interpreted as originating by globulation of lava under water from a very small vent or from a hole in the crust of a flow. The rods, spherules, etc. are composed of dark brown glass with rare amygdules and a few chloritised olivine crystals. The sparse interstitial material consists of chlorite and very pale green fibrous aggregate of tremolite (?).

(d) Zoned Lavas. South of City of Melbourne Bay there are extensive out-crops of lavas ranging in thickness from 15 cm. to over one m. They characteristically show chilled upper and lower surfaces and the coarser amygdaloidal core is, in many cases, zoned in similar fashion to the example of Plate 16, No. 1. The chilled margin, 3-5 cm. thick, passes down to a highly vesicular

Plate 15 No. 1:- Shower droplet bed, 200 m. north of Conglomerate Creek.

Plate 15 No. 2:- Close up of Shower droplet bed, 200 m. north of Conglomerate Creek. The coin has a diameter of 2.5 cm.

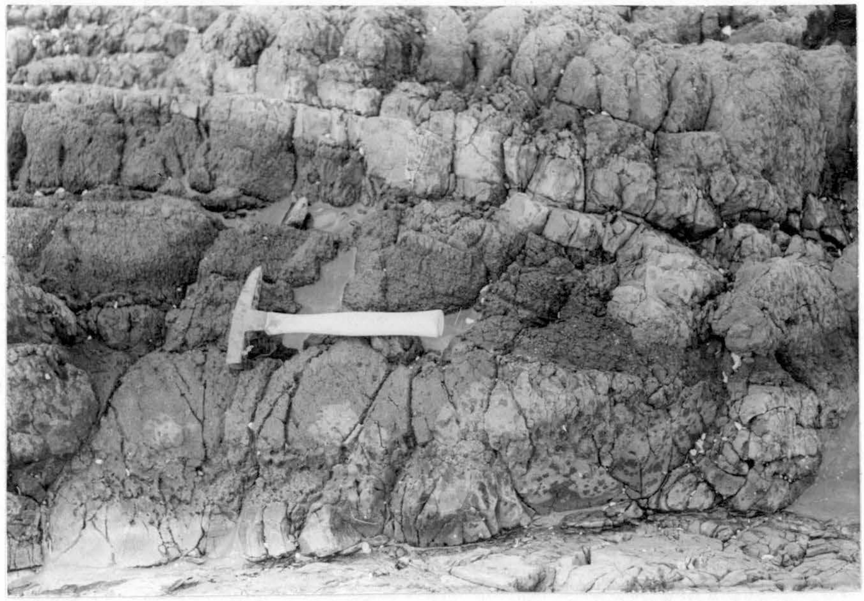


Plate 16 No. 1:- Zoned lava flow, 400 m. north of Conglomerate Creek.

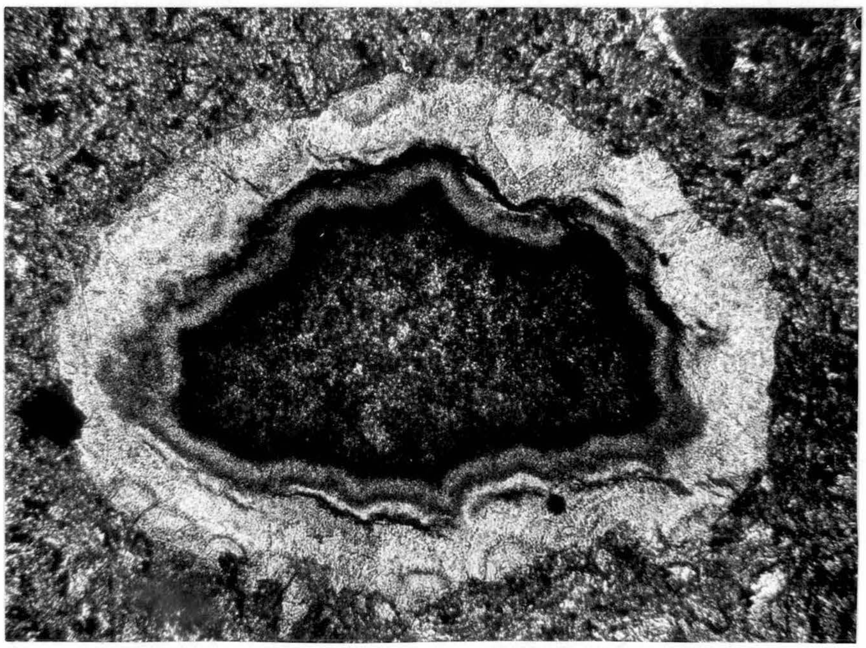


Plate 16 No. 2:- Colloform hydrogrossular in amygdule, same locality specimen 32425. X90.

zone which in turn overlies a less vesicular layer of similar composition (see also Scott, 1951a). The chilled margins consist of chloritised olivines and albite laths and a very fine-grained groundmass, the amygdaloidal zone of augite, chloritised olivine and albite and the third zone consists of coarser augite and albite crystals in a fine-grained groundmass. Amygdules are filled by chlorite, epidote, calcite, prehnite, albite, sericite, fine-grained quartz and to some extent hydrogrossular (see also Scott, 1951b). The hydrogrossular shows colloform texture in some amygdules (Plate 16, No. 2). These flows tend to vary in thickness in similar fashion to "necking" in tectonic boudinage (Plate 6, No. 1), and in places the upper and lower chilled margins actually come in contact, all the inner vesicular section being absent. This structure probably develops during flow and soon after the chilling of the margins, due to "stretching-out" of the thin, highly mobile, gas-charged core of the flow. These flows are not interbedded or associated with tuffs and appear to represent a phase of very quiet extrusion of gaseous lava with low viscosity. The surfaces of the flows are characterised by concentric "ropy" structure (Plate 17, No. 1).

These structures are confined to the thin surface skin and are relatively small scale, the maximum radius being about one m. They appear to have formed from the slow upwelling of very small streams

Plate 17 No. 1:- Concentric structures on surface of ropy lava, 400 m. north of Conglomerate Creek.

Plate 17 No. 2:- Amygdaloidal flows, 300 m. south of Conglomerate Creek. Photo by A.H. Spry.

of lava through the chilled crust, the lava stream oozing out slowly from its hole in a series of concentric ripples that were soon chilled. There is no sign of the large-scale twisting and contortion of the pitch-like lava shown by the pahoehoe flows but Scott (1951a) has pointed out that in many respects the flows resemble pahoehoe lava.

(e) Amygdaloidal Lavas. These are seen 400 m. north of Conglomerate Creek, overlying the zoned lavas just described. At first sight they appear to be poorly bedded tuffs (Plate 17, No. 2) but in fact consist of flows 10 or more cm. thick and composed of highly amygdaloidal material. Thin sections (32426) show that irregularly shaped amygdules filled by chlorite occupy 60-70% of the rock, the interstitial material containing olivine pseudomorphed by chlorite, and also tiny granules or feather-like crystallites of clino-pyroxene (?) and feldspar (?). These thin flows appear to have been poured out as extremely mobile froth that formed a skin over the pre-existing floor, the flows probably succeeding one another in rapid succession.

(f) Massive Lavas. These appear mainly at the base of the volcanic succession, also within the mudstone-dolomite sequence in Cumberland Creek, and near the top of the succession near Grassy.

The flows are several tens of metres thick and are crudely and randomly jointed. Patches up to a cu. metre or so are altered to epidote or epidote and albite, and veins of these minerals are common (31809, 31812).

The massive flows are typical augite-albite spilites with doleritic to basaltic textures (31815, 31808, 31813). Scott (1951a) has described unusual almost feathery, curved augite textures similar to features described in New Zealand spilites (the "cervicorn" augites of Battey, 1956) and attributed to very free diffusion during crystallisation (Walker, 1952). This is presumably related to the low viscosity. The augite occurs within albite in ophitic texture and the possibility of the albite being primary has been discussed by Scott.

Near Grassy, some depressions in massive flow surfaces are occupied by lenses of reddish-brown, finely-banded tuff that possibly flaked off the flow crust. This reddish tuff consists of a very fine, almost granular aggregate of feldspathic material, epidote and glass. South of Cumberland Creek, bands of yellow-green, compact tuff appear to have dragged by movement of an overlying massive spilite flow.

Summary of Lava Types on King Island

There appear to be two important types of lava on King Island - the augite-bearing rocks and the "olivine-rich" picritic rocks. The first type forms the type 1 pillows and the massive flows (and also dykes) and the second type forms the block lavas and the pillow and flow breccias. The zoned flows are mainly augite-type but the amygdaloidal are picritic.

The augite lava forms discrete flows and pillows that develop a tough, leathery skin while the picritic lava tends to break up and form fragmental rocks. Generally the picritic type has few amygdules though the amygdaloidal flows, which only occur in one small area, are an exception. The pic rite rocks have high water contents but do not appear to have given up gas by effervescence. This may be due to rapid chilling (and it is noted that they are poorly crystallised) but the water content may have been derived from deuteric reaction with volcanic gases and sea water. Scott (1951a) has drawn attention to similarities between some of the King Island flows and the aa and pahoehoe flows of Hawaii. According to Macdonald (1953) the aa lava is less vesicular, more viscous, and at a lower temperature than pahoehoe lava, comparisions which could very well fit the picrite and augite types of King Island respectively. The pahoehoe lavas also forms a tough skin while the aa does not. However, the Hawaiian lavas are of practically the same composition, one flow may show both aa and pahoehoe features at various stages of its eruption, and the aa lava is better crystallised than pahoehoe. In these respects, they are unlike the King Island volcanics and, of course, there is a lack of pillow lavas, and breccias and tuffs such as are found at King Island.

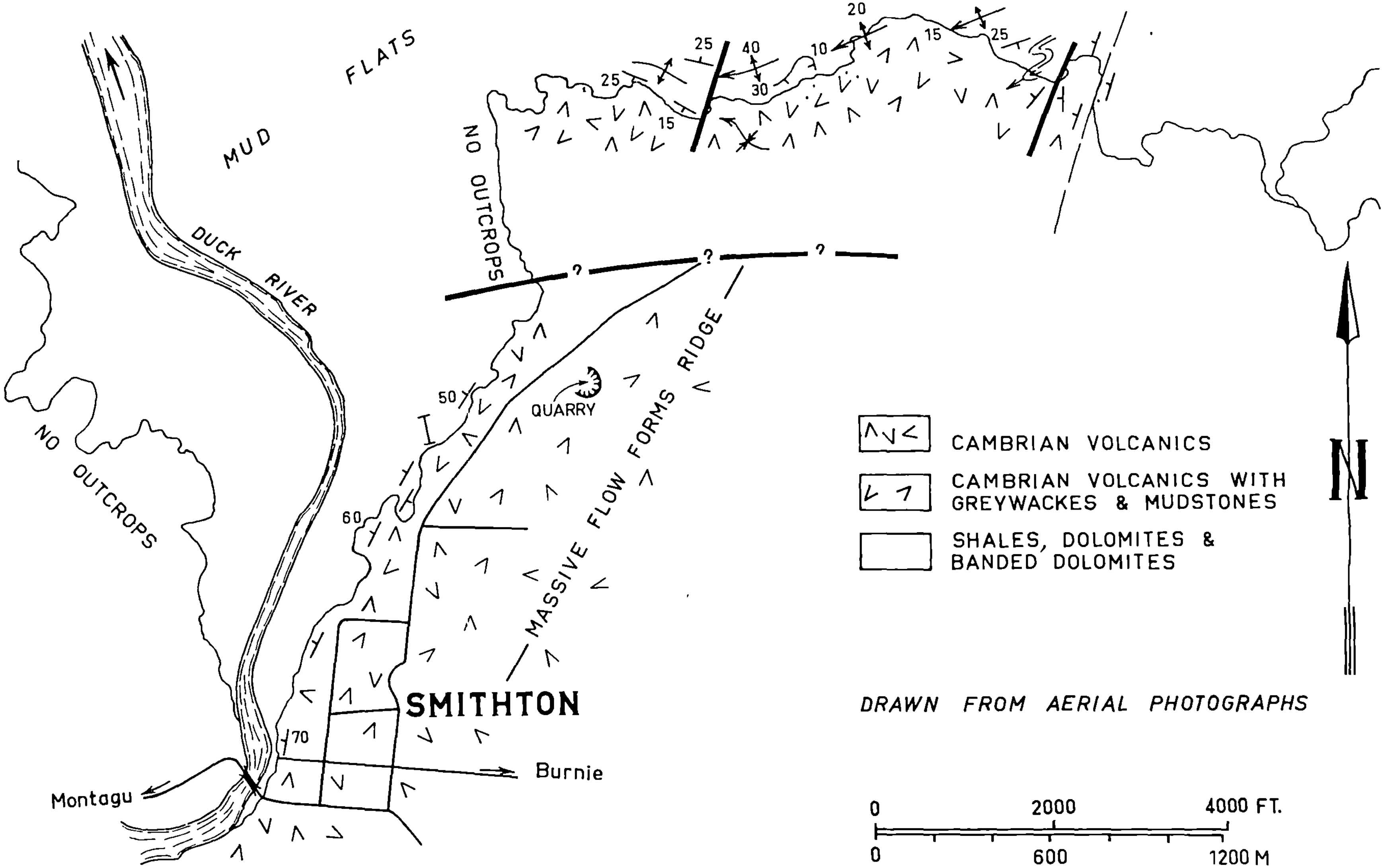
Other West Coast Occurrences

No additional types of flow are found among the remaining spilite occurrences, most of which are not well exposed

Crudely developed pillow lavas also occur over a small area near Penguin (Scott, 1952) and at Smithton (Carey and Scott, 1952) but have not been found elsewhere; most of the Penguin spilite are massive. Zoned lavas occur between the Spero and Mainwaring Rivers but have not been examined in detail. The lavas there are interbedded with thin tuff bands and shales and vary from about 15 cm. to about 30 m. thick. Brecciated lavas are particularly common and they appear to be autoclastic in type. A similar rock occurs on Lynch Creek (Scott, 1954; Solomon, 1960) and consists of irregularly shaped masses of spilite up to several cm. diameter within a spilite matrix. The breccias probably formed by fragmentation of chilled skins during movement of hotter underlying fluid.

In the Smithton area (Fig. 6) flows vary from a few cm. to about 100 m. thick; they are all massive and featureless and are interbedded with mudstones, greywackes and tuffs. They have been described by Nye, Finucane and Blake (1934) and Carey and Scott (1952), the latter authors being the first to identify the rocks as spilite flows. An unusual feature of the rocks is the abundance of relatively coarse-grained doleritic or pegmatitic textures and flows a few feet thick may be composed of albite laths up to 3 mm. long (e.g. 32440). They are all rich in hematite and form a rather distinctive tholeiitic group within the spilitic suite. In the road-metal quarry ½ mile north of Smithton the spilite (32449, 32435, 32439, 32441) is particularly rich in amygdules of quartz (and some chlorite) up to 3-4 cm. diameter. The quartz is present as a fine-grained, almost fibrous aggregate and as crude spherulites. Quartz also occurs interstitially in the groundmass. The spilite from this quarry is veined by quartz and epidote and some of the quartz veins contain a black lustrous mineral ringed by malachite and azurite. It is amorphous, non-radioactive and largely copper and iron (by X-ray spectrograph).

Fig. 6: Geological sketch-map of the Smithton area, showing the distribution of spilitic volcanics.


MUD FLATS
DUCK RIVER
NO OUTCROPS
NO OUTCROPS
QUARRY
MASSIVE FLOW FORMS RIDGE
SMITHTON
Montagu
Burnie
CAMBRIAN VOLCANICS
CAMBRIAN VOLCANICS WITH GREYWACKES & MUDSTONES
SHALES, DOLOMITES & BANDED DOLOMITES
N
DRAWN FROM AERIAL PHOTOGRAPHS
0 2000 4000 FT.
0 600 1200 M

The thickness of the volcanic succession is difficult to determine because of folding (see Figure 6) but must be over 300 m.

Tholeiitic spilites (32452) also occur in Cambrian mudstones ¼ mile east of Montagu which is 10 miles west of Smithton.

The spilites from Zeehan (the Montana Melaphyre), Penguin and Curtain Davis have been described by Scott (1954, 1952a). A distinctive feature of the Montana Melaphyre is the presence of glass in the groundmass and the presence of spherulitic quartz rock in some of the lavas.

Quartz spherulites occur in rocks at Magnet which were identified as spilitic lavas by Scott (1954). As these rocks are unusual and had been interpreted previously as intrusive (the "Magnet Dyke") they will be discussed later in some detail.

Mineralogy of the Spilites

The principal mineralogical features have been admirably described by Scott (1951a, b, 1954) and only a few features (e.g. the feldspar optics) require enlargment. Albite and augite and/or chlorite form most of the rocks and are usually the sole phenocrysts in porphyritic texture or the major components in intersertal (or doleritic) texture.

In 32905, from near Queenstown, ophitic texture between augite and cloudy albite laths is well displayed. This example is clearer than similar textures described in the King Island spilites by Scott (1951b). Variolitic textures and plumose arrangements of feathery albite and augite are also present. Scott figured an albite-augite intergrowth that is almost myrmekitic in form and appears to be unique to spilites. Though initially uncertain (1951a, b) as to the primary or secondary origin of the albite, Scott later (1954) concluded that it is secondary after an earlier plagioclase. The remainder of the spilites (groundmass or mesostasis) is composed of iron oxides (generally magnetite) with variable but small amounts of sericite (from feldspars), calcite, epidote, quartz and some ilmenite and sphene.

The albite generally occurs as elongate or stumpy to subhedral laths up to 2 mm. long or as thin microlites. It occurs mainly in porphyritic or doleritic texture. Most crystals are clouded by fine-grained alteration products or inclusions of micas and chlorite. Replacement by secondary minerals such as sericite, calcite, epidote, chlorite and some quartz is common. In some rocks (e.g. 32865, Lynch Creek) the feldspars are completely replaced by relatively coarse aggregates of sericite.

The most common twin laws are albite and Carlsbad, and albite-Carlsbad and pericline twins have also been observed. The feldspars of the spilites (and other igneous rocks) were investigated by the universal stage technique of Turner (1947) but using more recent data incorporated in new curves by Slemmons (1962). Because of the fine grain and high degree of alteration, determination of only a few twin axis orientations was possible and these indicated a composition near An_0. Extinction angle measurements (for albite twins) on the universal stage show a composition range from An_0 to An_5. The few twin axis measurements plot near the low temperature curve and several 2V determinations show values from 82° to 90° positive, indicating low temperature or near-low temperature optics. The feldspars appear to be similar to those in the keratophyres, for which more numerous determinations were possible because of the coarser grain-size.

Water-clear albite occurs in late veinlets, in amygdules and in the groundmass as fine-grained, irregular shaped aggregates. In all these cases it appears to be formed later than the majority of the rock. It is not observed replacing the earlier altered albite except where cross-cutting veinlets appear to be of replacement type, though in 32865 there is some indication that it replaces feldspar and also pyroxene. Clear albite is relatively rare in the spilites and is much more common in the keratophyres. In some cloudy albite crystals, small irregular patches of clear albite may be seen, but it is not clear whether this is an exsolution or a replacement phenomenon. In all specimens collected by the author albite is the only feldspar present in the spilites, an observation similar to that made by Scott. However, slide 636, collected by J. Molline from the West Magnet Mine, has a coarse doleritic texture with almost unaltered labradorite (An_{58}) laths up to 2 mm. long, and interstitial pools occupied by pale green chlorite, calcite and magnetite (?). It is prominently amygdaloidal with chlorite, calcite, and quartz spherulites. It is markedly different from any Tertiary basalts in Tasmania and is similar to spilites apart from its feldspar composition; however, there must remain some doubt as to its age or its field occurrence. If part of the Cambrian spilitic suite, it is the only indication of a parental, more calcic feldspar.

Augite is present as granules and occasionally as euhedral crystals. It is generally fresh (a feature remarked upon by Scott, 1954) and occurs as phenocrysts, interstitial material to feldspar networks, or in ophitic texture. Where little augite is present, the percentage of chlorite is high. Scott (1954) has, by chemical analysis, identified the augite from the Lynch Creek lavas as diopsidic and of the salite variey (with 21% Ca0). The optical properties fall in the range expected from the chemical composition (Deer et al., 1962).

Similar optical properties have been found in augite from other spilites and also in augites from gabbroic rocks (e.g. 31737A).

The augite is altered in varying, but generally slight, degrees to chlorite and at least some of the occasional ragged hornblende crystals in 32922 (Lynch Creek) appear to be derived from breakdown of augite.

Scott (1951a) tentatively identified pigeonite in the groundmass of her aa lavas but the writer has been unable to confirm this.

Olivine crystals pseudomorphed by pale green chlorite occur in several lava types on King Island but no olivine mineral remains.

Chlorite occurs as (a) an alteration product of augite; (b) as inclusions (?) in feldspar; (c) as interstitial or groundmass material of apparently primary origin; (d) as clots and veinlike segregations up to several sq. cm. in area; and (e) in amygdules with albite, calcite, etc. Strong colouration is typical of the rocks with plentiful chlorite. Carey and Scott (1952) referred to a bright green chlorite in the Smithton spilitic rocks and they suspected it was a nickel-rich variety, though without any chemical data. Scott (1951) suggested chlorite in breccias (aquagene tuffs ?) on King Island is also nickel-rich but again without confirmation.

In most cases, the chlorite in thin section is medium to pale green, may be slightly pleochroic and in most cases, has anomalous Berlin Blue interference colours. Scott measured β values of 1.592 and 1.626. Chlorite that is almost colourless in thin section is rare except in the spilites at Zeehan and Magnet, in which it is common as an amygdule filling; this chlorite has strong Berlin Blue or brown interference colours and a β of 1.594, the lowest figure obtained by the writer among the chlorites. Chlorite collected from amygdules in the Montana Melaphyre was analysed by the X-ray method described in the section on the Lyell Schists and shown to have the generalised composition 7.3 Mg 4.7 Fe.

Calcite (and siderite ?) and epidote are patchy in distribution and are present in crudely devebped masses scattered through the groundmass, and also in amygdules and veinlets. They are in many cases associated with clear albite.

Actinolite occurs rarely in veins up to several cm. width, e.g. at Lynch Creek and the Henty River.

Hydrogrossular occurs as amygdule fillings, in veinlets, as replacements of phenocrysts and in the mesostasis in some of the King Island spilites (Scott, 1951a, 1951b, 1954) and also occurs replacing glass in the aquagene tuffs. It is clearly a late stage mineral. It also occurs in a volcanic breccia (646) at the head of the Arthur River south-west of Waratah (Plate 18). The breccia consists largely

Plate 18 :- Hydrogrossular (top centre) in auto-clastic explosion-breccia, Waratah. X95.

of feldspar crystals, quartz spherulites and basaltic fragments up to 1 mm. in size in a variable matrix of chlorite, iron oxides and calcite. Part of the matrix consists of irregular veins and microbotryoidal hydrogrossular. The breccia is probably an autoclastic explosion-breccia (Wright and Bowes, 1963) formed by gaseous explosions of semi-consolidated spilitic lava.

The cell-size of the hydrogrossular given by Scott (1951a, p. 127) seems rather high for the relatively low percentage of water shown in her analysis. A measurement of the cell-size in hydrogrossular from specimen 5175 by the writer gave $\underline{a}$ = 12.00.

Quartz occurs in amygdules, in the mesostasis of the tholeiitic types at Smithton, and in lenticular veins and segregations. At Penguin, fine-grained quartz and jasper lenses are relatively common (Plate 19, No. 1) and are in some cases associated with epidote and jasper.

Tremolite, as a very fine-grained, felty aggregate, forms much of the groundmass of several of the King Island lavas and is probably a late magmatic ~~alteration~~ product.

Pyrite is a very minor but ubiquitous constituent of the spilites; concentrations occur in the amygdaloidal zones of type 1 pillows.

Native copper is more common as specks up to a few mm.

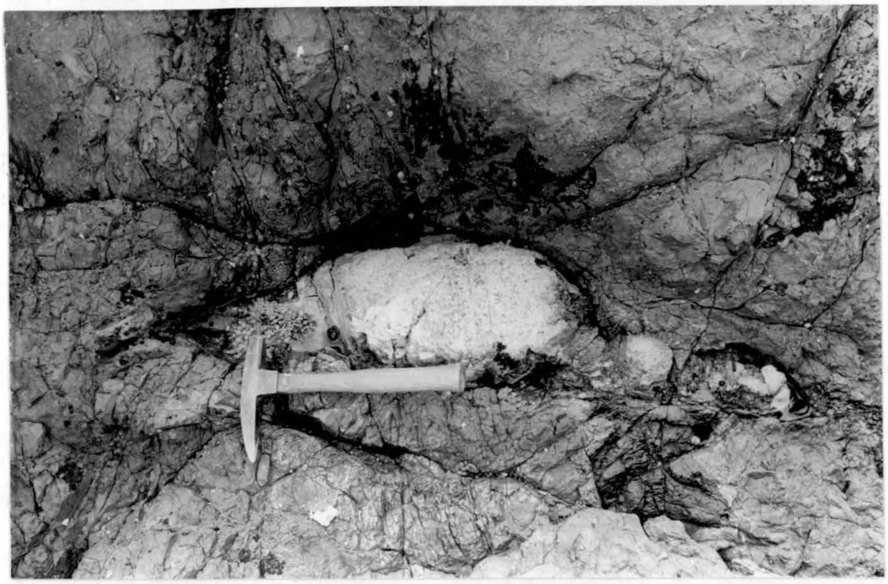


Plate 19 No. 1:- Lens of finegrained, yellowish quartz in massive spilite, half a mile east of Penguin.

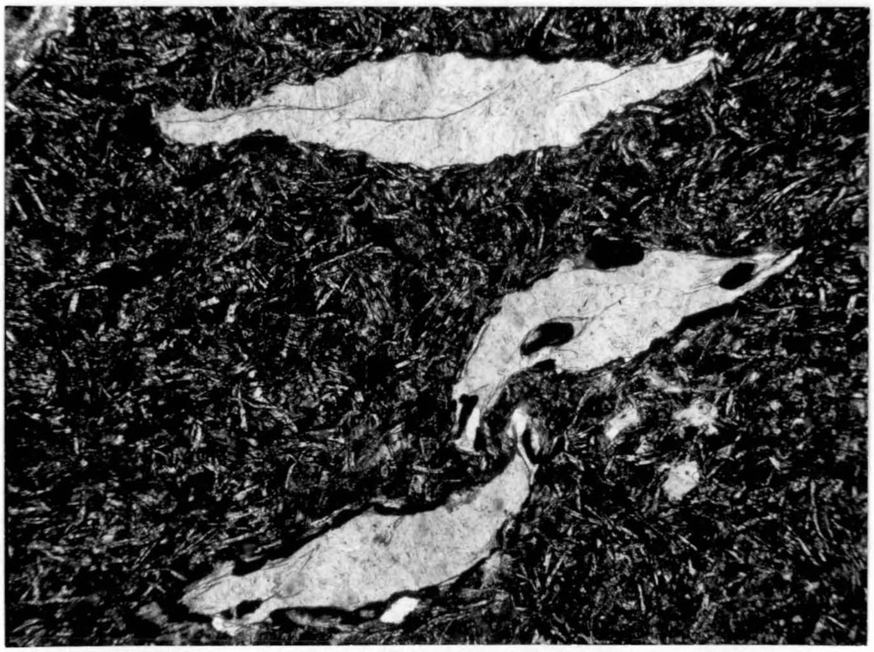


Plate 19 No. 2:- Amygdules of chlorite in the Montana Melaphyre. X40.

diameter in the groundmass of most spilite flows. It has not been found in amygdules and has not been found concentrated in veinlets or segregations.

Amygdules are rarely absent, are ovoid or spherical in shape and may be filled with chlorite (most commonly), clear albite, calcite, epidote, prehnite, hydrogrossular, quartz, clinozoisite and iron oxides. A common zonal arrangement shows an outer rim of quartz, an inner rim of chlorite and a core of calcite. Epidote prisms may form an outer rim or occupy the core. Parts of the flows, not necessarily the upper portions, are composed largely of amygdules e.g. much of the Montana Melaphyre consists of chlorite and chlorite-calcite ovoids in a partly devitrified base (Plate 19, No. 2). A feature of the Zeehan amygdules is their ovoid shape and parallel arrangement with some (e.g. 30096) having curled ends. These features are probably due to flow, though in most cases the groundmass microlites or crystals do not show parallel orientation. Many of the chlorite amygdules in the Montana Melaphyre have a rim of granular quartz (e.g. 256, 4848 - Plate 20) and a few have chlorite rims and are filled largely by chalcedonic silica. In several amygdules in spilites from the Waratah area, the chlorite filling is rimmed by a thin zone of quartz spherulites, each spherule being 0.1 mm. or so in diameter (Plate 21, Nos. 1 and 2). In others

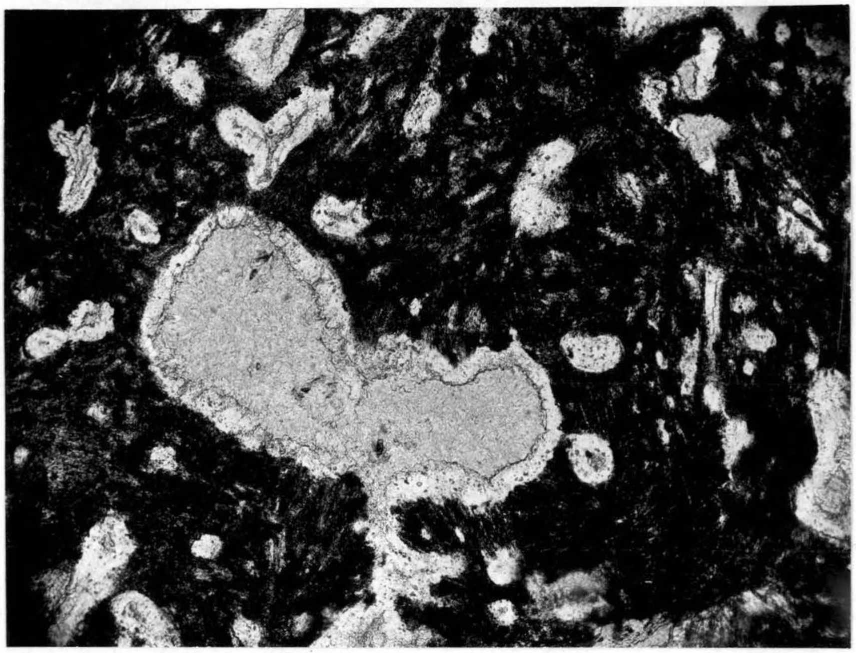


Plate 20 :- Rims of quartz to chlorite amygdules in the Montana Melaphyre. X95.

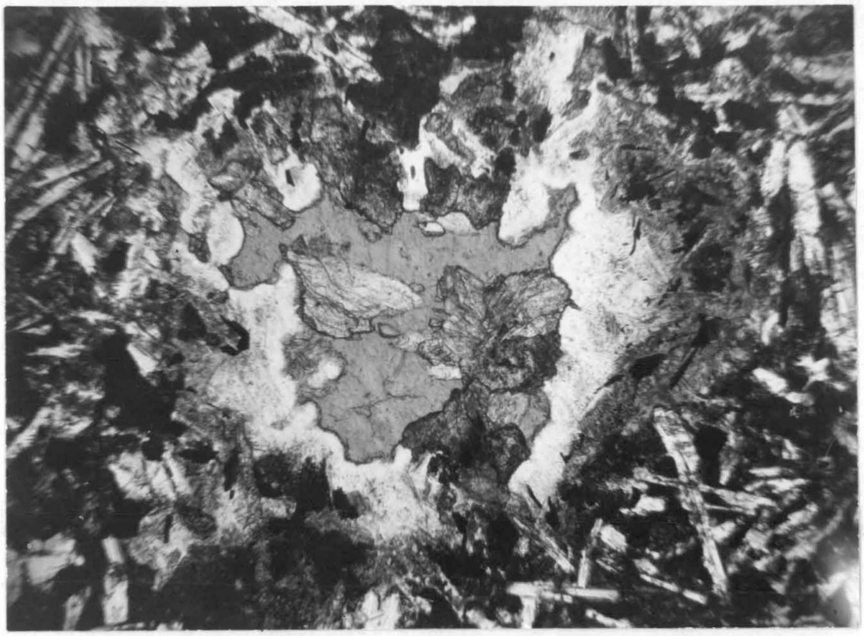


Plate 21 No. 1:- Quartz spherulite rims to chlorite amygdules in spilite from near Waratah ordinary light. X36.

Plate 21 No. 2:- Quartz spherulite rims to chloritic amygdules in spilite from near Waratah, crossed nicols.

(30662) almost the entire amygdule is composed of spherulites and the groundmass may consist in whole or in part of similar material (Plate 22, Nos. 1 and 2). A rock made up almost entirely of quartz in spherulitic form occurs within spilites near Magnet and the Montana Mine (Zeehan) and is described on page 67 . The origin and age of this material is one of the major problems of the Tasmanian lavas and as the host rocks of the Magnet orebody are involved in this, they are described in a separate section.

Glass in the Spilitic Rocks

Spilite from Lynch Creek (32865) has hyalopilitic texture with up to 15% of dark, iron-rich glass in the groundmass (Plate 23, No. 1). The glass contains skeletal microlites of feldspar and irregular, somewhat hazy masses of fine-grained albite-aggregate that appear to have formed by devitrification of the glassy material. Glass is an important constituent of the Montana Melaphyre and many specimens show in thin section a fine-grained network of feldspar microlites with ferruginous, "dusty" glass and partly devitrified, weakly birefringent material (4848, 30096, 50G96-102).

The presence of glass in these rocks has been commented on by Carey (1954, p. 112-113), who has pointed out that a discontinuity

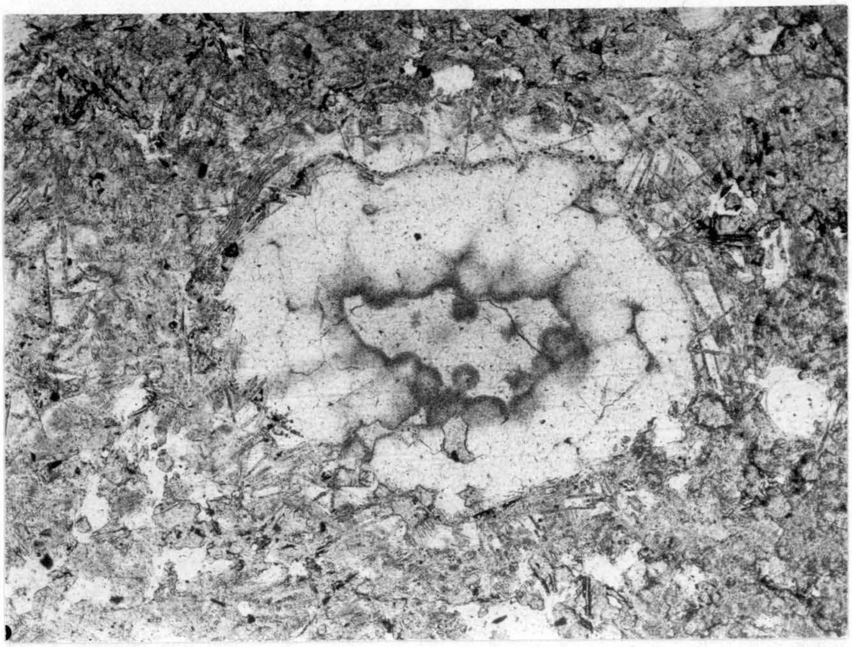


Plate 22 No. 1:- Quartz spherulite-chlorite amygdules from near Waratah, ordinary light. X70.

Plate 22 No. 2:- Quartz spherulite-chlorite amygdules from near Waratah, crossed nicols. X70.

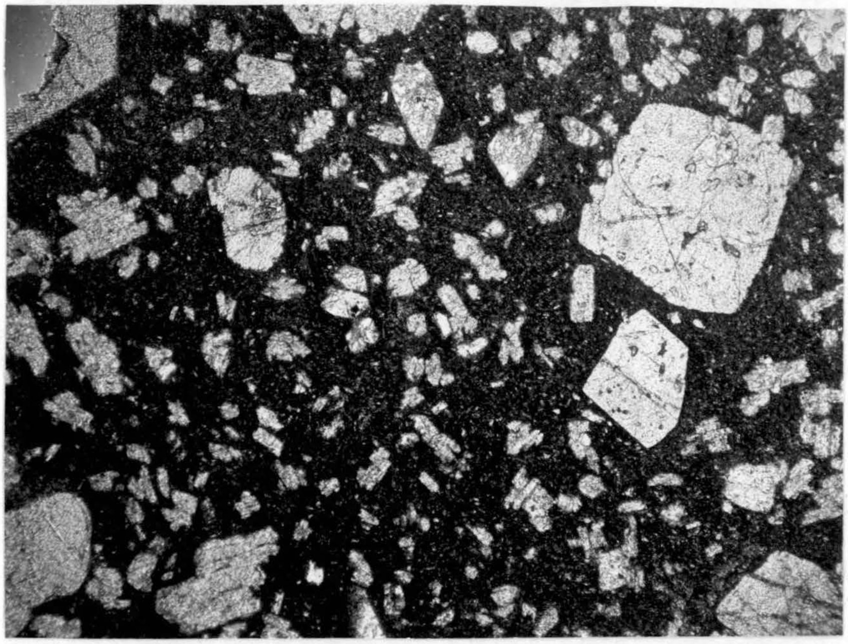


Plate 23 No. 1:- Glassy matrix to spilite from Lynch Creek, Queenstown. X22.

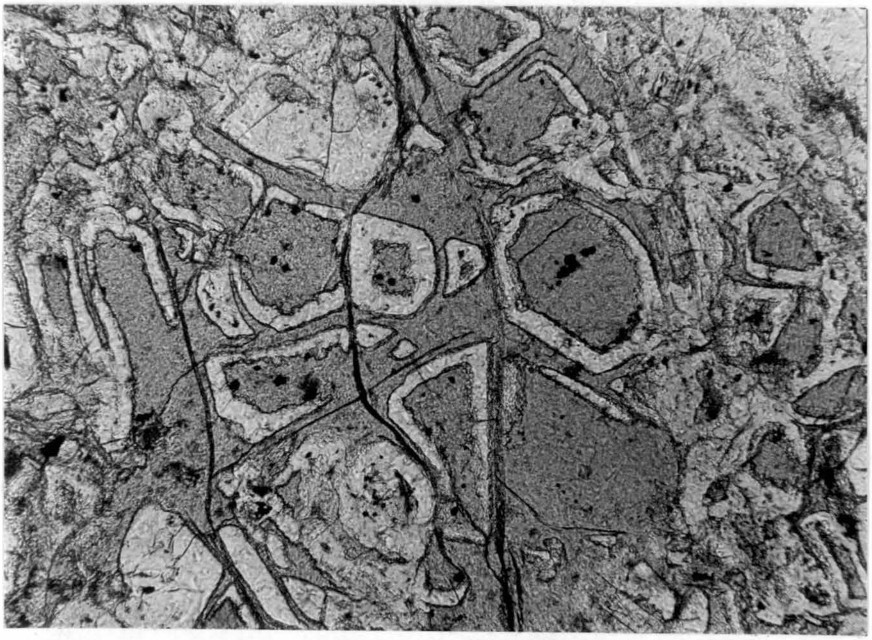


Plate 23 No. 2:- Hieroglyphic texture in variolitic spilite (31831) from the Magnet Mine. X85.

exists in the properties of glass-forming substances in the region of the "glass transformation temperature" (T_g), which is the temperature at which the viscosity of the glass is of the order of 10^{13-14} poises (Pryde and Jones, 1952). At and above T_g the viscosity falls rapidly with rising temperature. If, after sudden quenching, the glass is never heated to the vicinity of T_g, it may take viscous glasses (e.g. basaltic) a very long time, say, 10^{17} seconds ($3,000 \times 10^6$ years) to devitrify.

Marshall (1961) has also pointed out that thermal devitrification (due to temperature changes) is very slow and that an average glass would devitrify in 500×10^6 years (the age of the spilites ?) only if held at about 225°C. However, in the presence of water, devitrification is much faster and is very sensitive to temperature. Hence the presence of glass at Zeehan and Lynch Creek suggests that since eruption these lavas never reached a temperature of say, 200°C, and that water had little or no access to the glass.

The vitrophyres of King Island differ in being largely devitrified. These rocks probably cooled slowly and in the presence of water, and devitrification may have been very rapid. The viscous deformation of the glass particles indicates a temperature above, say, 450°C and the presence of hydrogrossular indicates a temperature between 300°C and 700°C (Yoder, 1950). Marshall's figures (e.g. p. 1515) indicate that devitrification might well have been very far advanced in a few years under conditions of slow cooling from 300 to 400°C. The devitrification, then, quite probably took place during a "volcanic-hydrothermal" stage, along with the development of chlorite, hydrogrossular and other secondary minerals.

The "Magnet Dyke"

A long, narrow strip of igneous rocks occurs at the base of the Cambrian (?) sediments from the Magnet Mine to the Persic Mine, a distance of about 5 miles (Fig. 7). This mass of igneous rock has occasioned some controversy, Twelvetrees (1900 b) and Nye (1923) considering it to be a complex dyke, and Scott (1954) suggesting on petrological evidence that it represents a strip of volcanic rocks. The "Dyke" contains several lead-zinc fissure lodes, the largest of which is at the Magnet Mine.

The "Dyke" is generally some 60 to 100 m. in width, reaching its maximum thickness at the Magnet Mine where it is some 300 m. wide. At the mine it has a west dip and is more or less conformable (Fig. 8) and in outcrop it is essentially parallel with, and close to, the base of the Cambrian (?) sediments (Fig. 7). It consists over

Fig. 7: Geological map of the Magnet-Mt. Bischoff area. Geology by M. Solomon and D.I. Groves. See also Solomon (in press, Appendix A paper no. 3).

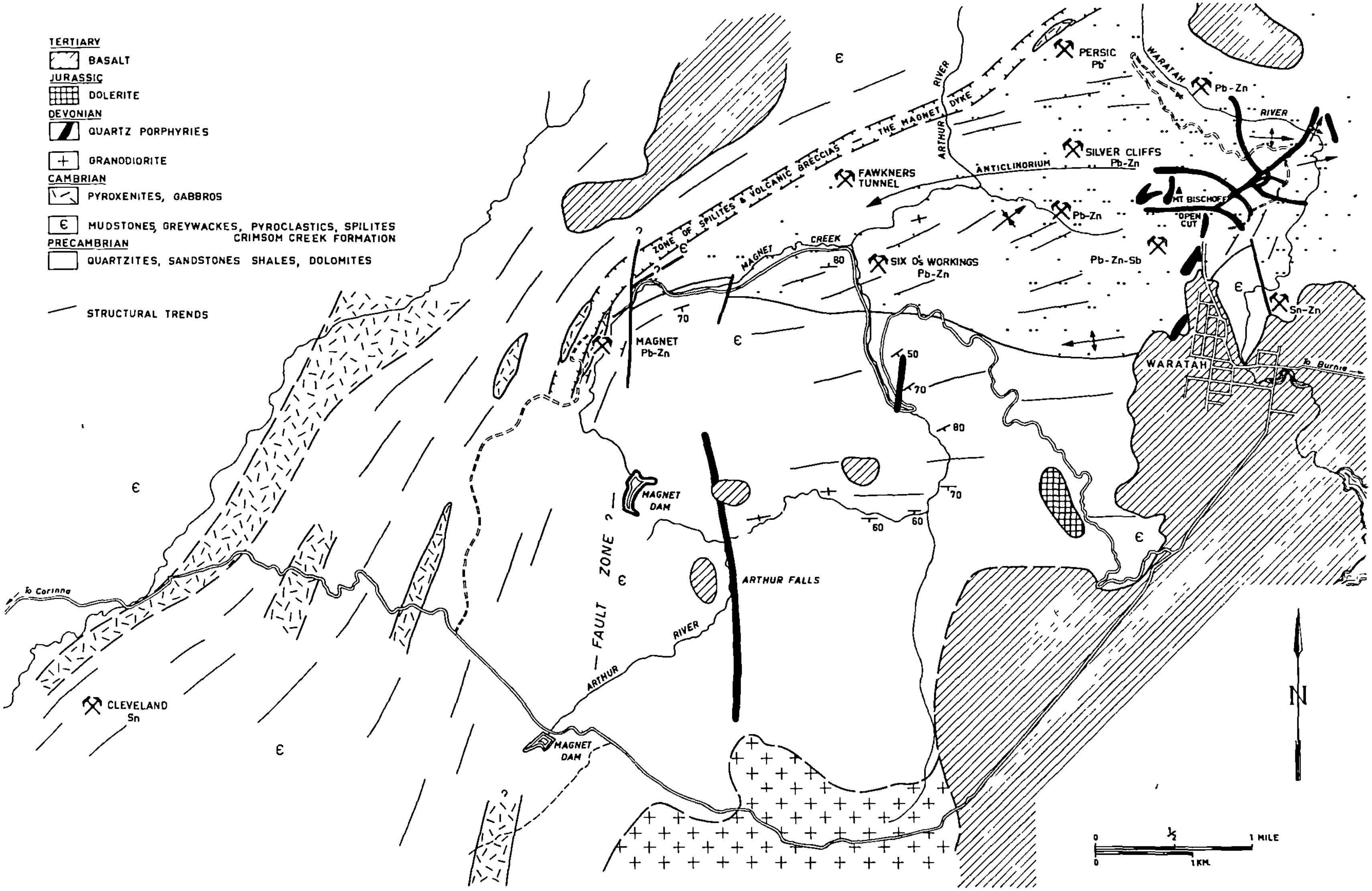
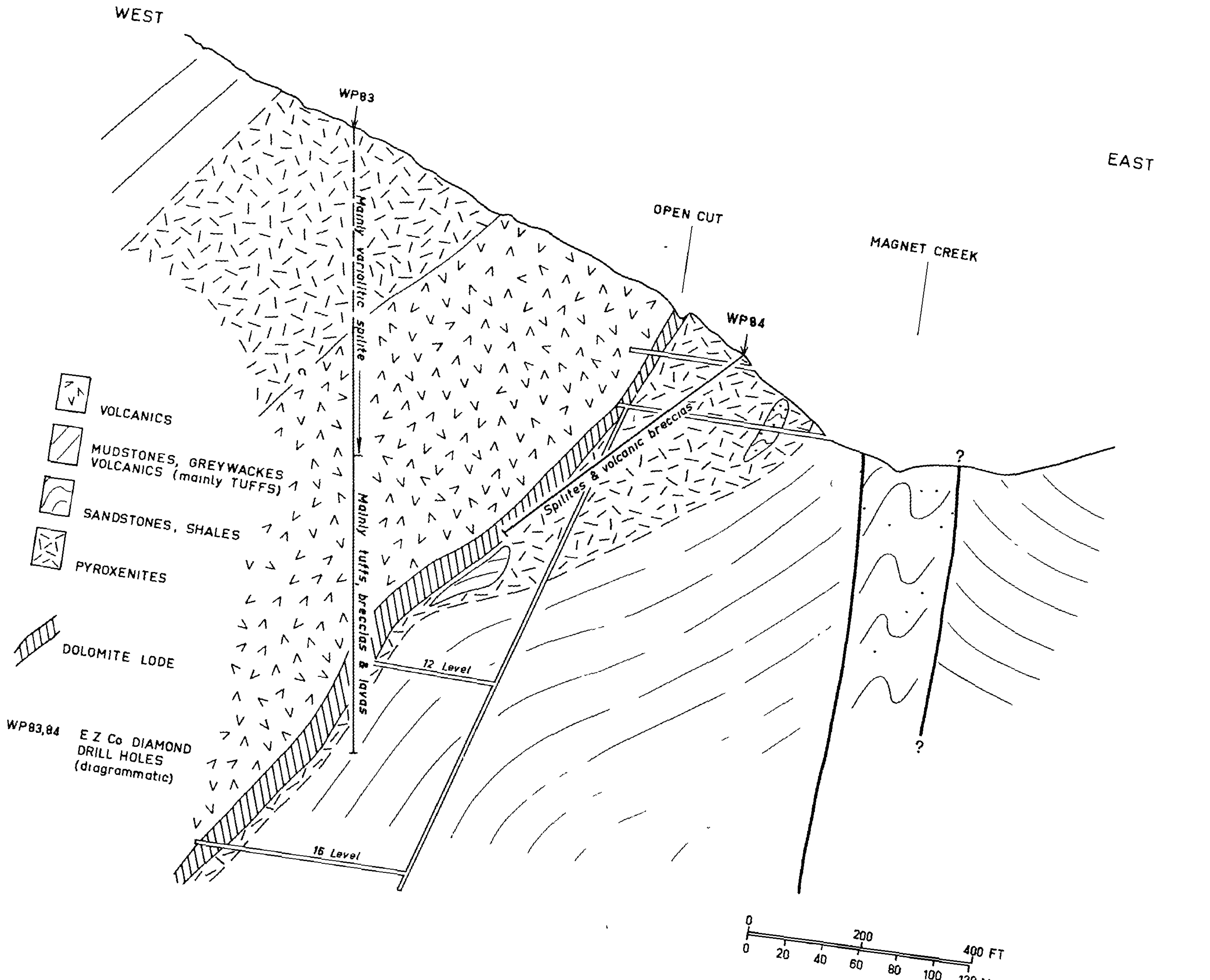

TERTIARY
BASALT
JURASSIC
DOLERITE
DEVONIAN
QUARTZ PORPHYRIES
GRANODIORITE
CAMBRIAN
PYROXENITES, GABBROS
MUDSTONES, GREYWACKES, PYROCLASTICS, SPILITES
CRIMSOM CREEK FORMATION
PRECAMBRIAN
QUARTZITES, SANDSTONES SHALES, DOLOMITES
STRUCTURAL TRENDS
ZONE OF SPILITES & VOLCANIC BRECCIAS
THE MAGNET DYKE
ARTHUR RIVER
PERSIC
Pb
WARATAH
Pb-Zn
RIVER
SILVER CLIFFS
Pb-Zn
FAWKNERS
TUNNEL
ANTICLINORIUM
MT BISCHOFF
OPEN CUT
Pb-Zn
Pb-Zn-Sb
MAGNET CREEK
SIX O's WORKINGS
Pb-Zn
Sn-Zn
MAGNET
Pb-Zn
WARATAH
To Burnie
MAGNET DAM
FAULT ZONE
ARTHUR FALLS
ARTHUR RIVER
To Corinna
CLEVELAND
Sn
MAGNET DAM
N
0
½
1 MILE
0
1 KM.

Fig. 8: Cross section of the Magnet Mine area, showing Electrolytic Zinc Co. drill holes projected about 60 m. from the south. Partly compiled from Twelvetrees (1900b) and Nye (1923).



WEST
EAST
WP83
WP84
OPEN CUT
MAGNET CREEK
Mainly variolitic spilite
Mainly tuffs, breccias & lavas
Spilites & volcanic breccias
12 Level
16 Level
?
?
VOLCANICS
MUDSTONES, GREYWACKES
VOLCANICS (mainly TUFFS)
SANDSTONES, SHALES
PYROXENITES
DOLOMITE LODE
WP83,84
E Z Co DIAMOND
DRILL HOLES
(diagrammatic)
0
200
400 FT
0
20
40
60
80
100
120 M

most of its length of rock which has been previously named a diabase porphyrite. At the Magnet Mine, however, the section is more complex, Twelvetrees (1900b) describing the "Dyke" from east to west as websterite porphyrite, diabase porphyrite, and orbicular or spheroidal websterite (Figure 8).

The "websterites" on either flank of the diabase appear to be identical mineralogically, and spheroidal textures have been found in one place on the east side of the diabase, increasing the similarity (Nye, 1923). The rock is extensively altered, even to considerable depths below the surface, and almost all of the primary minerals have been converted to chlorite, talc, etc. However, the alteration appears to have been by volume for volume replacement and the crystallographic outlines of the essential minerals are in many places preserved. Unfortunately, specimens of these rocks are rare and most of the information presented is derived largely from old slides in the University of Tasmania and Department of Mines collections. Thin section 701 shows euhedral outlines of orthopyroxene-like mineral up to 2 mm. in length and arranged at random. They now consist almost entirely of talc (X-ray identification), generally in fairly coarse plates aligned parallel to the orthopyroxene (?) axes so that the mineral outlines exhibit straight and symmetrical extinction. The matrix consists of sheaves of talc plus chlorite and quartz. The talc is sufficiently coarse to obtain very clear interference figures which proved to be uniaxial positive; apparently the 2V for this talc is zero.

The orbicular websterite contains almost spherical bodies from a few tenths of a millimetre to 30-40 cm. in diameter. The spheres are identical with the host rock throughout and their origin is unknown. Thin section 637 is from a rock almost identical with 701 but which contains spherical aggregates a few mm. across of talc, chlorite and quartz with dark rims. It also contains rounded aggregates of quartz and sericite (?) that look much like recrystallised sandstone and the rock has a distinctly fragmental appearance. A sample of orbicular material from the Launceston Museum (1957:34:14) consists of prominent altered pyroxene (?) laths up to 5 x 2 mm. in section, arranged randomly in a yellowish matrix. In thin section the pyroxenes (?), which constitute 70% of the rock, are replaced by talc and some chlorite (X-ray determination) and locally by a fine-grained quartz aggregate. The mesostasis is largely chlorite and quartz. There are also irregular masses of very fine-grained altered rock showing perlitic (?) cracks and in places this rock also has a fragmental appearance.

In thin section 709, pyroxene (?) outlines are visible but the entire rock is replaced by quartz and carbonate. The quartz pseudomorphing the pyroxene (?) is scaly and the scales are aligned so that they parallel crystallographic outlines of the earlier mineral. This prefered orientation of the quartz individuals emphasises the random arrangement of the pyroxene (?) crystals.

The term websterite was applied to these rocks by Rosenbusch (in Twelvetrees, 1900b) because he believed they originally contained enstatite (or bronzite) and diallage in approximately equal quantities, similar to a rock near Webster, Carolina, containing hypersthene and diallage (Johannsen, 1938, p. 460). The presence of pyroxenes is deduced only from the crystal outlines and type of alteration but these criteria are probably sufficient to class the rocks as pyroxene-rich. It is suggested they be referred to as altered pyroxenites (or pyroxenite breccias for the fragmental types).

The pyroxenites are lenticular and have been proved to thin out north and south of the Magnet Mine and also down dip (see Nye, 1923 and recent drilling) but they reappear at the North Magnet and Persic mines, on the footwall of the "dykes". The pyroxenites are concordant and are presumably sill-like. Their intrusive character is supported by the presence of lenses of Bischoff sediments and Crimson Creek rocks within the footwall pyroxenite because the lenses are unlikely to have reached their position by tectonic movement. The fragmental rocks in the footwall mass may have formed during the intrusion, the fragments consisting of local sediments and volcanics. Possibly the presence of the pyroxenites is related to the extreme alteration seen in the diabase porphyrite.

Drilling by the Electrolytic Zinc Co. (Rosebery) through the diabase porphyrite has provided an excellent opportunity to examine the "Dyke"; two holes were drilled several hundred feet south of the Magnet Mine, apparently beyond the southern limit of the pyroxenites (Figure 7). The core of these holes has recently been made available to the writer and has been examined in some detail. As a result it is cle~r that the diabase porphyrite consists of a conformable succession of volcanic rocks of spilitic composition. The log of hole WP 84 is given below:-

0-62 ft: no core

62-363 ft: Dark grey-green rock, mainly composed of grey spherical bodies, ½-1 cm. in diameter, which abut against each other or coalesce to form reniform masses (32809, 32810). These are variolites and consist of radiating, feathery, and "bow-tie" masses of very fine albite and chlorite containing smaller chlorite-quartz amygdules. In places (e.g. near 145 ft.) they have a thin (1 mm.) rim of carbonate. Intervening areas are dark green and are filled by chlorite and laths of albite. X-ray diffractograms confirm the presence of albite, chlorite and calcite in this rock. Veinlets of calcite are common, particularly between 312 and 325 ft.

363-375 ft: Greenish grey, fine-grained amydaloidal spilite (32811) with patches of variolitic rock as described above. Some of this rock is fairly pale grey and closely resembles parts of the Montana Melaphyre. The amydules are not greater than 2 mm. diameter and are composed of calcite and/or chlorite. The remainder of the rock is almost entirely composed of albite laths with interstitial chlorite, calcite, and iron oxides.

375-572 ft: Richly variolitic spilitic rock (32812); 60-70% consists of roughly spherical variolites from a few mm. to 1.5 cm. diameter and composed of albite, plus quartz and chlorite. The albite occurs in feathery sheaves and "bow-tie" masses, similar to those between 62 and 363 ft., and the quartz largely in circular, amygdule-like bodies. Long, thin laths of faintly greenish material appear to be chloritised incipient crystals. Some variolites have calcite rims.

Interstitial material consists of quartz and chlorite with a few albite laths. In 31831 ~~(collected by the author)~~ the quartz occurs as crescents and rings in chlorite (Plate 23, No. 2) to form "hieroglyphic" texture. This rather strange texture probably represents a mass of amygdules that have been pressed together in places to produce cross sections ranging from circular to almost square.

Analyses of "altered variolites" quoted by Nye (1923, p. 48) and given in Table 4 are of rocks that are probably similar to the quartzose variolites just described.

572-950 ft., and 963-1,098 ft: These sections are very variable but are characterised by well defined banding in grey-green tuffs and breccias of spilitic type (32814, 32815) together with some spilite lavas (e.g. at 584 ft. - 32820). Typical of the coarser rocks is a breccia between 1,040 and 1,041 ft. (32817). It consists of fragments (possibly bombs ?) up to 9 cm. across of highly amygdaloidal spilite in coarse-grained, banded, spilitic tuff. Other smaller and very irregularly shaped fragments are much paler in colour and

Table 4

	1	2	3
SiO_2	73.00	60.00	99.11
Al_2O_3	9.95	11.97	Nil
Fe_2O_3	5.29	7.68	0.75
FeO	0.39	0.39	Tr.
MgO	3.84	6.50	Nil
CaO	2.30	4.35	"
CO_2	5.90	8.90	"
FeS_2	0.57	0.24	"
Total	101.24	100.03	99.86

1 and 2. "Variolite", same specimen, some carbonate removed, south adit, Magnet Mine (Nye, 1923, p. 48).

3. Spherulitic quartz rock, Montana Flat (Scott, 1954, p. 139).

may be keratophyric. Thin sections of fine-grained breccia (574-575 ft. - 32819) consist of angular fragments of fine-grained basaltic lava and quartzose albite porphyry. This porphyry consists of albite laths up to 0.3 mm. long with much smaller quartz crystals in a confused groundmass of quartz and albite; several of the albite phenocrysts are replaced by calcite and chlorite. Clearly some keratophyric rocks of fairly acid nature were present in the area at this time. Finely banded tuffs (e.g. at 1,056 ft. and 655 ft. - 32818) are conspicuous, and in many places show bands a few cm. thick of fine-grained spilitic breccia. The banded tuffs are in part very quartz-rich (e.g. at 1,032 ft. - 32816) and may be tuffaceous mudstones. Parts of this section are dark green and greasy and consist entirely of talc (32813) which appears to occur as "beds" within tuffaceous sequences. Another type of alteration is conspicuous through most of these rocks and consists of dolomite replacement and veining (dolomite determined by X-ray diffraction). The veinlets are so numerous and of such diverse orientations that in places little of the original rock is visible, even in thin section (e.g. at 834 ft. - 32821).

950-963 ft: This section is lode rock, consisting of white dolomite and a little chlorite and quartz with particles of sphalerite, and probably represents the southern extension of the Magnet lode.

Hole WP 83 (see Figure 8) shows that the footwall pyroxenite also thins out rapidly south of the mine, its place being taken by altered volcanics. The log is as follows:-

0-about 70 ft: dark green rocks, completely altered to aggregates of chlorite, quartz and calcite with patches of talc (32468, 32469).

70-102 ft: Volcanic breccias, with angular fragments up to 5 cm. diameter of spilite, mudstone and sandstone or tuff in a dark green chloritic matrix (32469).

102-126 ft: Green, fine-grained spilite with 1-2 mm. chlorite-filled amygdules (32470).

126-144 ft: Brecciated spherulitic spilite (32471).

144-154 ft: Reddish grey, coarse mudstone without banding and containing nodules and veinlets of pyrite.

154-173 ft: Fine-grained spilite laced with dolomite veinlets.

173-452 ft: Volcanic breccia with many fragments (up to 10 cm. across) of spilite, sheaved and laced by contorted dolomite veinlets. Some veinlets have a pale green colouration due to chlorite.

452-527 ft: Dark green chloritic rock veined intensively by sphalerite-dolomite; main mineralization at 452-454 ft.

527-561 ft: Fewer dolomite veinlets in green, chloritic (and talcose ?) sheaved volcanic rock.

Several thin sections of the "diabase porphyrite" from the University of Tasmania and Mines Department collections help to amplify the descriptions given above (Mines Department numbers contain N or G).

Nos. 35G4 and 35N6 are of feldspathic greywacke and mudstone respectively, these rocks being similar to those found elsewhere in the district and apparently forming part of the "Dyke". Nos. 35N8 (labelled "websterite") and 35N22 are basaltic breccias composed of fragments up to 5 mm. of fine-grained spilite, augite and feldspar in a quartzose matrix.

Slide 35N20 is fairly typical spilite with doleritic texture. It consists of long (0.5 mm.), thin laths of albite, granular and feathery augite and interstitial magnetite, green chlorite and a few ragged plates of hornblende. 35N15 is similar but the only ferromagnesian mineral is chlorite. Calcite is also present in interstitial pools and with chlorite and quartz, fills amygdules. Rosenbusch (in Nye, 1923, p. 47) described spherulitic quartz in amygdules.

35G10 is porphyritic, containing albite laths up to 3 x 4 mm. in a fine-grained doleritic base. This consists of albite, augite crystals, green chlorite and magnetite. The augite displays ophitic texture with albite and is partly altered to dusty brown material. The chlorite does not appear to be derived from the augite and is presumably of primary origin. Both 35N15 and 20 show trachytic texture in places.

The feldspars in the rocks just described vary from almost fresh to severely altered (sericitised ?). The process is carried to the extreme in 35N12 in which the albite lalths are entirely composed of a fine-grained sericite (?) aggregate; interstitial pools (and a few amygdules) are entirely calcite.

The remainder of the slides are largely extremely modified forms of spilite. 35G3 contains many small laths replaced by chlorite and 35G8 is largely composed of laths up to several mm. long now completely replaced by chlorite. It is suspected that the earlier mineral was albite, though some cross-sections resemble olivine. The matrix of 35G8 consists entirely of calcite and quartz, with amygdules composed of quartz spherulites and chlorite.

Thin section 710 is described as "silicified melaphyre" but consists of a single variolite of about 2 cm. diameter. The variolite consists of feathery albite sheaves, and irregular patches of green chlorite and clear quartz. Some of the quartz occurs as aggregates replacing feldspar (?) laths. Rosenbusch (in Nye, 1923, p. 47) also described quartz replacing feldspar and quartz encloses, and also replaces, albite in 35N19, another example of the variolitic rock. 35N16 is described as a variolite and presumably has variolites within it. It consists of an indistinct quartz aggregate that shows sheaf texture and indefinite extinction. It may once have been glassy, for within it are the forms of numerous brown, acicular crystallites, a feature also seen in the Montana Melaphyre and common in glassy lavas.

Thin sections 708, 35N10 and 35N13 are composed of quartz and chlorite, partly in crude aggregate, partly in hieroglyphic texture, and parts of the slides consist largely of quartz spherulites. Outlines of albite (?) laths are clearly visible in ordinary light and are now mainly quartz.

From these descriptions and drill hole logs, it is evident that the diabase porphyrite of the "Dyke" is actually a succession of spilites (mainly variolitic and altered) overlying a succession with volcanic breccias, tuffs and mudstones and rare spilites, similar to Cambrian rocks outside the "Dyke". The pyroxenites, however, must surely be intrusive and are discussed in a later section.

A feature of nearly all the Magnet lavas is the extensive "alteration", particularly by silicification. The abundance of silica is further emphasised by the presence in the lavas of spherulitic quartz rock, which also occurs at Zeehan in the Montana Melaphyre.

Spherulitic Quartz Rock

Attention was first drawn to this rock by Twelvetrees and Ward (1910, p. 19-20) who described the Zeehan material as follows:- "Macroscopically, it is somewhat variable in appearance. Some specimens are dull-white or greenish-white, and others have a prevalent reddish tinge. There is a white base, which has the appearance of porcellanite, and in this are set the spherules. The spherules are at times so closely packed that no base appears between them. They are pale-green or blood-red in colour, and in many cases both colours are exhibited by one spherule. The radial structure of the spherules is visible to the naked eye, and their average size lies between one-sixth and one-quarter inch in diameter.

"In thin sections the spherules are seen to be cloudy with inclusions, some containing moving bubbles. Their radial nature become apparent between crossed nicols. A dark cross then appears, the arms of which do not move with the rotation of the slide, but remain parallel with the vibration-planes of the nicols. At times the cross has double arms. The spherules become deformed when closely packed together, yet they preserve, their radial structure. Their optical character is positive, as would be the case whether they were composed of quartz or microfelsite. In general appearance the material is quite homogeneous, and resembles quartz.

"Only one mass of this rock has been found. It was situated in the Montana Flat, near the mine buildings, and has now been broken up and removed, for on account of its handsome appearance when cut and polished the rock attracted no little attention. It is situated in a locality where the immediately adjacent rocks are slate and spilite breccia."

The fine-grained groundmass containing the spherules is chalcedony but in one specimen (5029) this forms only a thick white rim around the spherules, the remaining areas being occupied by a grey, coarse-grained, quartz aggregate.

Rocks rich in quartz spherulites occur west of Waratah in small bodies within spilitic flows. One (30662b) consists of euhedral phenocrysts up to 2 x 2 mm. pseudomorphed by chlorite and lying in a "groundmass" of chlorite and quartz spherulites up to 0.2 mm. diameter. Lavas from the Waratah area containing amygdules partly composed of spherulites have already been mentioned and quartz forms entire amygdules (as coarse sheaves, e.g. 32449 from Smithton) or rims to amygdules, as in parts of the Montana Melaphyre (e.g. 4848 from the Trial Harbour road). Both Rosenbusch (in Nye, 1923) and Scott (1954, p. 138) have described silicification of feldspars (replaced by quartz aggregate) to varying degrees in lavas from the Magnet ~~area~~ and Zeehan areas and several examples have been given in preceding pages. Partly silicified feldspars occur in tuffaceous sandstones west of Waratah and these sandstones also contain amygdule-like bodies of chlorite, and quartz and albite aggregates.

Twelvetrees and Ward (1910) and Scott (1954) analysed the quartz rock from Zeehan and found it to be over 90% silica (Table 4). Scott assumed the quartz rock was the result of regional metasomatism of late Cambrian age and she described in some detail a transition from fine-grained quartz aggregate in the groundmass of the melaphyre to a rock composed entirely of quartz spherules. This description was based on study of a number of individual specimens apparently showing different stages of growth. The mechanism was supposed to be one of solid-liquid metasomatism and the chemical changes involved were illustrated by graphs showing reduction in everything but silica. The quartzose variolites from Magnet were thought to represent stages in the metasomatism of the basaltic rocks.

Twelvetrees and Ward regarded the rock as "highly altered", though previously (1900a) Twelvetrees had suggested it was an offshoot from a granite mass. Rosenbusch, describing the rock at Magnet, suggested it was an altered variolite.

The writer suggests that the rock is of deuteric origin, for the following reasons:-

(a) The quartzose rock has only been noted near Waratah and Zeehan and occurs only in, or close to, spilitic rocks. This distribution hardly seems compatible with regional metasomatism.

(b) Rims of quartz spherulites in amygdules suggest that their formation is related to the amygdule phase and not to some later metasomatism.

(c) Siliceous pockets and veinlets are typical of the Tertiary basalts of Tasmania and radial spherulitic bodies have been recorded in basaltic rocks from many localities (e.g. Bryan, 1962).

(d) Silicification of feldspars appears to be related to the quartz-spherulite rock, and silicified feldspars occur as detrital (or pyroclastic ?) grains in Cambrian sediments apparently interbedded with the lavas near Waratah. This suggests that the silicification was more or less contemporaneous with vulcanism.

(e) The metasomatic transition described by Scott is not necessarily valid, based as it is on a carefully chosen collection of specimens, and the different examples may equally well mark different forms and phases of crystallisation from a siliceous solution.

It is suggested that the quartz rock represents residual siliceous solutions derived from crystallisation of spilite, the siliceous residues segregating by diffusion, by injection into the lava, or by concentration in gas channels. The first stage of their formation is the segregation of quantities of siliceous material. At some critical stage of temperature and saturation, with the lava possibly still mobile, crystallisation commences within the solution at a number of irregularly distributed points and proceeds outwards to form radial crystallites - these spherical bodies so produced have no "skin" and may merge or coalesce. Bryan (1954, 1962), discussing the origin of variolites in Queensland tachylyte, suggests that the nucleation agencies for this type of crystallisation may be minute steam bubbles. This process may continue until the entire rock is composed of spherulites or it may be arrested by a change of conditions, causing the remaining melt to crystallise between the spherulites as chalcedony or quartz.

Variations from the typical quartz spherulite rock occur when the siliceous solution contains crystals of earlier minerals such as feldspar or ferromagnesian mineral, producing porphyritic rocks with quartz spherulites in the groundmass.

There is a possibility that the spherulites have formed by devitrification of glass but this seems unlikely because of the presence of spherulites in amygdules and the lack of spherulites in partially crystallised glass within the adjacent lavas.

The occurrence of quartz spherulites in amygdules, as previously described, calls for fucher discussion. The postulated history of the majority of the amygdules is described below. While the lava is still molten but after crystals of feldspar and perhaps ferromagnesian minerals have commenced to grow, bubbles of gas form and start to rise through the lava, keeping their shape because the gas pressure within is approximately the same as the radial pressure exerted on the bubble. The bubbles receive additional material by impact with other bubbles and by distillation of gases of lower vapour pressure from the magma. Once this two-phase assembly is present, the gas-liquid interface becomes a source of attraction for material that is capable of lowering the surface tension at this interface. Ringwood (1956) and Stansfield (1928) have shown that sulphides particularly (because of their low surface tension) tend to move to such interfaces and it is interesting to note that galena and pyrite are quite common in some acid amygdaloidal lavas of Cambrian age in Tasmania. Molecular liquids, such as water, have low surface tension and silica may also have low surface tension.

Bubbles may also collect small crystals by adhesion and these may remain attached until solidification. As cooling proceeds the gas in the bubble condenses and a pressure gradient is therefore set up; this will control the flow of mobile material towards the bubble. The bubble will not collapse because the cooling host melt is now so viscous that it traps the bubble and at the same time supports its walls. Crystallisation from solution will proceed from the rims of the bubble inwards and will continue as long as sufficient feed is maintained; changes in the feed, differences in solubility, etc. may produce a zonal structure. Thus the crystallisation proceeds inwards for amygdules and out from a centre in the spherulites. It is also possible that some crystallisation within the amygdule may proceed by diffusion through the semi-permeable "skin" that must form around the amygdule late in its history. Diffusion from outside the "skin" towards the amygdule may result in growth of an outer rim that makes room for itself by enlarging the amygdule or squeezing the "core". Such growth might account for the highly irregular surface seen in many of the amygdules. Thus the core material or an amygdule may not necessarily be the last phase to form.

The quartz-spherulite rims to chlorite amygdules in slide 636 (Magnet area) might well have developed in the manner just describedbut why numerous spherulites should form in preference to normal crystals is not known. Presumably the reason is bound up with the speed at which crystallisation is induced, deposition from numerous centres achieving a quicker solidification than growth via larger crystals.

To summarise, it is suggested that the quartz spherulitic rock is a localised product of the late stages of lava solidification and it is not the product of regional metasomatism.

Alteration of the Spilites

It is clear from the preceding descriptions that most of the spilites and associated rocks have been considerably altered, with the production of typical "secondary" minerals such as tremolite, chlorite, epidote, hydrogrossular, quartz, talc, sericite etc. The mode of occurrence of many of these minerals suggests that they formed as late magmatic minerals; the presence of chlorite, epidote, hydrogrossular and quartz in amygdules and of concentric tremolite veinlets in pillow cores suggests a magmatic origin and similar conclusions have been derived from a study of the quartz spherulite rock. The jasper clots and veinlets in the Penguin spilites are also probably primary. The dolomite veining of the Magnet volcanics and possibly also the heavy talc development may well be related to a later phase of mineralization and some of the minerals are typical products of very low-grade regional metamorphism, particularly albite, chlorite and epidote. However, hydrogrossular and tremolite imply a higher temperature, probably related to volcanic processes.

In 1951, Scott believed the alteration of the spilites on King Island to be caused by a nearby intrusion of Devonian granite but in 1954 she suggested the spilite alteration was related to the late Cambrian (Jukesian) orogeny.

Depositional Environment of the Spilites

Scott (1951a) believed that the spilites were subaerial, the pillow lavas developing where the flows passed into local bodies of surface water (an aqueous environment for the pillows seems essential). Carlisle (1963) has good evidence for believing that pillow breccias and aquagene tuffs also require aqueous conditions. He states (p. 68) that "the obvious fact that pillows settled in soft fossiliferous limey muds at the base, the gradational continuity from one rock type to the next, the abundant evidence of quenching, globulation, and pasty flow, and the laminated and graded tuffs with load casts from overlying breccias at the top of the section leave no doubt whatsoever that all of these rocks are of submarine origin". The significance of some of these items is uncertain but the remainder certainly point to a submarine origin for the British Columbia rocks. The pillow breccias (whole and broken pillows) described by Henderson (1953) at Yellowknife and in Iceland are associated with pillow lavas and are believed to be of subaqueous origin. Carlisle conceives lava of very low viscosity and relatively low gas pressure emitting into water and globulating "vigorously and promiscuously" to form isolated pillows that settle into tuff. Disaggregation of pillows to form breccias he thinks may be caused by slumping "essentially during extrusion" and disruption may have been aggravated by steam explosions. Similar processes operating in a similar environment probably produced the King Island breccias but it seems highly probable to the writer that in Tasmania extrusion took place in part into water-saturated tuffs and that elongation and deformation of pillows and thin flows took place during movement through the tuff-water medium under shallow cover. That flow has occurred in the host tuffs is indicated by a non-tectonic foliation that wraps around pillows and flows. The similarity of the King Island rocks to thoseof British Columbia, the presence of pillows and some bedded tuffs point to at least an aqueous origin for the volcanics.

The sediments enclosing the spilites at Smithton, Magnet and the Wanderer River area are greywackes and banded mudstones that are generally accepted as of marine origin, and as also are the sandstones and dolomitic rocks of the spilitic succession at Zeehan (Carey, 1953, Solomon, 1962).

Nomenclature of the Spilitic Lavas

Scott stated (1954, p. 147) that "the volcanics comprise a spilitic suite consisting of picrite basalts, spilites, keratophyres, albite dolerites and associated pyroclastics ....." The word spilite is one of the less exactly defined terms of igneous geology and has to used with caution. In a review of the characteristics of spilites, Vallance (1960), after summarising previous usage of the term, was forced to include all rocks called spilite by earlier workers. After taking this step, he finds that spilites have a wide range of chemical composition, the only significant features (in comparison with basalts) being (a) highly variable distribution of components

(b) high water content

(c) generally high carbon dioxide content

(d) generally high alkali content

(e) generally low lime content.

He is not prepared to accept high soda content as a characteristic of spilites. Mineralogically, however, Vallance has less difficulty in distinguishing spilites. They are characterised by albite (or possibly sodic oligoclase), rare quartz, lack of olivine, fresh pyroxene, chlorite, epidote, clinozoisite and less commonly, calcite.

Turner and Verhoogen (1960, p. 258-260) have described a similar mineralogy for spilites and also high Na_20, low K_20 and rather low Al_20_3; they also point out that these rocks are commonly associated with "normal" basalts. Distinct from these "general" definitions are those by Sundius in 1930 (rocks with normative feldspar $<$ 40%An) and Wells in 1922 (low Fe^{3+} / Fe^{2+}).

Apart from the unusual German and Swiss uses, there does seem to be common ground in the mineralogy and this perhaps should be the starting point for identification. It appears that this mineral assemblage shows a wide range in chemical composition and that gradations exist to basalts, picrite basalts, andesite and other lavas. Probably this is why Vallance (1960) includes among his spilites many marginal rocks; yet he does not include the spilites from Smithton because Scott called these albite dolerites. The variation in composition between flows, within flows (Battey, 1956) and even within pillows (Vallance, 1960; Vuagnat, 1949) makes concise definition of a spilite composition rather meaningless. Apart from the mineralogy there are other general features that assist in identifying spilites, particularly the associated igneous rocks and the depositional environment.

It is concluded that spilites are volcanic rocks characterised by albite (or possibly sodic oligoclase), chlorite, augite, epidote and less commonly calcite; quartz is rare and olivine absent. Their chemical compositions have basaltic affinities but compositions of the member of any spilitic assemblage show considerable variation; some may be rich in potash. They tend to be associated with basaltic rocks and a geosynclinal environment. The use of terms like albite-dolerite appears of little significance as this is only a textural variation and the writer has found in Tasmania, as have others working elsewhere, that doleritic and basaltic textures are distributed irregularly in groups of spilitic rocks and even within flows.

The writer would suggest that all the basic volcanics in West Tasmania are spilitic and may be referred to collectively as spilites. The intermediate and acid types are keratophyres, albite andesite etc., the whole collection forming a spilite-keratophyre association, suite or assemblage.

Chemical Composition of Tasmania Spilitic Rocks

With the previous discussion on nomenclature in mind, the Cambrian basaltic rocks of Western Tasmania may now be examined. They are clearly spilitic in mineralogy and, like other spilitic suites, include picrite basalts and rocks transitional between spilites and basalts (see Spry, 1962b, p. 257-259). One third of the analyses available (Table 5) show "basaltic" soda contents and high alumina, the only spilitic characteristic according to Turner and Verhoogen being low potash. Of Vallance's criteria the only common feature is the high water content as compared to basalts. Several of the analyses described as spilites by Scott (1952, Table IV) have all but one of the characters required of a spilite by Vallance, and Turner and Verhoogen. An interesting point is that most of the analyses have higher lime contents than basalts whereas Vallance claims spilites have lower lime contents. The Fe^{3+} / Fe^{2+} ratio is extraordinarily variable.

Table 5: Tasmanian Spilites

	1	2	3	4	5	6
SiO_2	46.53	48.35	52.61	47.40	48.24	50.01
TiO_2	0.21	0.78	0.72	0.29	0.70	0.73
Al_2O_3	10.51	16.82	13.03	19.19	17.55	15.38
Fe_2O_3	0.62	2.85	3.90	1.48	1.05	4.86
FeO	8.27	10.21	8.48	8.26	7.04	9.21
MgO	17.36	4.46	5.10	3.60	5.27	5.85
CaO	10.04	9.55	7.26	11.25	10.43	6.35
Na_2O	1.90	3.78	5.60	3.40	5.58	4.77
K_2O	0.22	0.42	0.42	1.31	0.97	0.40
H_2O+	3.71	2.32	1.65	3.32	2.88	2.60
H_2O-	0.31	0.32	0.10	0.34	0.17	0.23
MnO	0.16	0.10	0.19	0.13	0.12	0.21
P_2O_5	Tr.	n.dt.	Tr.	n.dt.	0.10	0.09
CO_2			0.05		0.11	0.13
S			0.08			
FeS_2						
Total	99.84	99.96	99.19	99.97	100.21	100.82
qu						
or	1.30	2.48	2.48	7.74	5.73	2.36
ab	16.08	31.14	46.19	18.16	19.13	40.36
an	19.50	27.69	9.18	33.23	19.98	19.37
ne		0.46	0.65	5.75	15.22	

Table 5: Tasmanian Spilites (continued)

		1	2	3	4	5	6
di		24.29	16.47	21.87	18.95	25.03	8.81
hy	en	0.33					3.65
	fer	0.12					3.03
ol	forst	23.97	5.35	4.93	3.64	4.52	5.99
	fay	8.93	8.12	4.93	6.14	4.22	5.47
mag		0.90	4.13	5.65	2.15	1.52	7.05
hem							
calc				0.11		0.25	0.30
ap						0.23	0.21
ilmen		0.40	1.48	1.37	0.55	1.33	1.39
Total		95.82	97.32	97.36	96.31	97.16	97.99

Table 5: Tasmanian Spilites

	7	8	9	10	11	12
SiO_2	44.20	45.00	46.64	46.88	50.16	53.20
TiO_2	1.44	0.32	0.54	1.48	2.00	1.74
Al_2O_3	20.80	24.64	23.45	19.04	18.01	19.15
Fe_2O_3	9.98	7.58	9.00	5.15	13.98	7.72
FeO	5.84	3.48	1.80	9.03	4.15	3.87
MgO	3.18	6.30	3.26	3.62	1.84	2.89
CaO	10.40	9.07	10.91	10.67	1.40	1.24
Na_2O	1.71	1.06	2.09	2.00	4.43	5.17
K_2O	0.36	0.46	0.50	0.55	0.83	0.58
H_2O+ / H_2O-	1.44	3.20	3.00	2.00	2.10	4.00
MnO	1.06			Tr.	Tr.	Tr.
P_2O_5	0.15	0.05	0.09	0.13	0.25	0.30
CO_2						
S		Tr.	0.27	0.04	Tr.	0.05
FeS_2						
Total	100.56	101.16	101.55	100.59	99.15	99.91
qu	5.94	8.39	4.67	3.07	16.18	14.79
or	2.13	2.72	2.95	3.25	4.90	3.43
ab	14.38	8.97	17.68	16.92	37.48	43.75
an	48.06	44.67	53.13	41.35	5.31	4.19

Table 5: Tasmanian Spilites (continued)

		7	8	9	10	11	12
cor			6.03			7.88	8.48
ne							
di		2.03		0.32	8.91		
hy	en	7.15	15.69	7.97	6.90	4.58	7.20
	fer	1.87			7.57		
ol	forst						
	fay						
mag		14.47	10.29	4.24	7.47	7.58	7.43
hem			0.48	6.08		8.75	2.60
calc							
ap		0.35	0.12	0.21	0.30	0.58	0.70
ilmen		2.73	0.61	1.03	2.81	3.80	3.30
Total		99.11	97.97	98.28	98.55	97.04	95.87

<u>Table 5: Tasmanian Spilites</u>

	13	14	15	16	17	18
SiO_2	48.20	46.72	47.6	50.6	43.34	47.98
TiO_2	2.36	0.48		0.5	3.15	0.99
Al_2O_3	14.74	18.25	17.7	20.3	15.69	17.89
Fe_2O_3	9.22	2.38	13.7	10.7	1.87	4.86
FeO	7.82	7.73	n.dt.	n.dt.	8.76	6.11
MgO	5.00	7.81	6.55	4.4	11.19	5.74
CaO	5.91	7.86	8.2	6.6	2.88	7.65
Na_2O	2.89	2.64	2.0	2.3	2.49	3.17
K_2O	0.44	1.32	1.3	1.8	0.67	0.74
H_2O+		0.11			6.00	
H_2O+ / H_2O-	} 2.50		} 1.5	1.9		} 2.97
H_2O-		4.43			0.39	
MnO	0.74	0.07	n.dt.	n.dt.	0.05	0.19 (15)
P_2O_5	0.385	n.dt.			0.93	0.25 (10)
CO_2			0.6	0.4	1.89	0.18
S	0.068					
FeS_2					0.30	
Total	100.273	99.80	100.25	100.5	99.60	
qu	9.45				6.12	
or	2.60	7.80			3.96	
ab	24.45	22.33			21.07	

Table 5: Tasmanian Spilites (continued)

		13	14	15	16	17
an		25.95	34.05			
cor						10.87
ne						
di		0.68	4.02			
hy	en	12.19	3.56			27.87
	fer	3.04	2.12			7.66
ol	forst		10.27			
	fay		6.73			
mag		13.37	3.45			2.71
hem						
calc						2.95
siderite						1.56
ap		0.90				2.17
ilmen		4.48	0.91			5.98
Total		97.11	95.24			92.92

Table 5: Tasmanian Spilites

1. Picritic spilite, King Island (Scott, 1952b, p. 87).

2, 3. Augite pillow lava, King Island (Scott, 1952b, p. 87).

4, 5. Massive spilite, King Island (Scott, 1952b, p. 87).

6. Spilite ("basaltic type") of Scott (1952b, p. 87).

7, 8, 9, 10, 11, 12. Tholeiitic spilites, Smithton (Nye, Finucane and Blake, 1934, p. 68 69 and 72).

13. Pegmatitic phase (?) of Smithton spilite; the "coarse gabbro" of Nye, Finucane and Blake (1934, p. 68).

14. Spilite, Lynch Creek, near Queenstown (Scott, 1952b, p. 87).

15. Spilite, Lynch Creek, near Queenstown (Solomon, 1960).

16. Spilite, " " , near Miners Ridge (Solomon, 1960).

17. Montana Melaphyre, Zeehan, specimen 5055. Analyst, Department of Mines, Tasmania.

18. Average of analyses 1 to 17.

The only other notable feature of a group of the Tasmanian lavas is the low titania content compared to many spilites and most basalts. Scott points out that the Al_2O_3 and TiO_2 contents are more in keeping with andesitic magmas though she does not suggest that the rocks are differentiated from andesite. She notes their similarity to the Porphyritic Central type basalt of Mull, a type which is also characteristic of orogenic provinces such as the San Juan area of Colorado and the Cascade Range, N.W. United States (Turner and Verhoogen, 1960) and in which andesites and basalts are closely associated. The presence of andesites in Tasmania increases the similarity.

The spilites in Tasmania may be divided into three types: the picritic, augitic and tholeiitic. The picrites are (or were) olivine rich, and low in alkalies and Fe_2O_3; the augitic types have no olivine, are better crystallised, and higher in alkalies and Fe_2O_3; and the tholeiitic types from Smithton have modal and normative quartz and more oxidised iron. The Montana Melaphyre (Table 5, No. 17) is tholeiitic in composition, though its textural features differ from thoseof the Smithton lavas. The picritic and augitic types are alkaline basalts.

Yoder and Tilley (1962, p. 352-3) have suggested there is a continuum between basalt types but indicate five groups ranging from oversaturated tholeiite with normative quartz to alkali basalt with normative nepheline. The tholeiites of Tasmania correspond to their over-saturated tholeiites and the remainder range across the other four groups, viz: saturated tholeiites, olivine tholeiites, olivine basalts and alkali basalts. Spry (1962b, p. 263) suggested the spilites have been derived from tholeiitic olivine basalt by differentiation and demonstrated the similarity of sodic differentiates developed during crystallisation of Jurassic dolerite to the average Tasmanian spilite. He suggested from a study of the SiO_2 v mafic index diagram (Fig. 15) that the principal trend of differentiation was iron enrichment and that the picritic types developed by olivine accumulation from an olivine basalt. The analyses show trends combining alkali, tholeiitic and Skaergaard properties. The Na-K-Ca and A F M diagram (Fig. 19) also show the spilites covering a wide spread in the alkaline, tholeiitic and calc-alkaline zones.

Both tholeiitic and alkaline affinities are revealed in the MgOvAl_2O_3 / SiO_2 diagram (Fig. 20) used by Murata (1960) and Eaton and Murata (1960) for the Hawaiian lavas. These authors believed the Hawaiian lavas are derived from a tholeiitic magma by fractional crystallisation of clinopyroxene but this suggestion has been strongly criticised by Yoder and Tilley (1962, p. 415). On the alkali v SiO_2 curves (Fig. 18) used by Rittmann (1962) the spilites are very similar to his alkaline (or Atlantic) rocks such as the Lower Palaeozoic volcanics of Scotland and the lavas of Hawaii. Finally, on the scheme proposed by Simpson (1954), involving femic and mafic indices (Fig. 16) the trend in iron enrichment is again apparent but the erratic distribution of the felsic ratio causes a spread that makes correlation with classic types impossible.

Available trace element information is given in Table 6. Co, Ni and Cr are higher in the picritic spilites compared to the augite types and Li, Sc and Ga are lower. All the elements but Zn and Pb fall within the range of various basalt provinces, between which there are considerable variations. There are unusually high concentrations of Zn, even though the X-ray analyses may be too high (an optical spectrograph analysis by the Australian Mineral Development Laboratories of Zn on 31658 gave 150 ppm Zn).

The single Pb analysis is also high for basaltic rocks in general. X-ray spectrographic analyses of the spilites gave substantially higher figures for Pb but these are not shown because they are not comparable with optical spectrograph and spectrophotometric analyses.

The only other trace element analyses of spilites known to the writer are by Amstutz (1953). His figures show unusually high

Table 6: Trace Elements (in p.p.m.) in Tasmanian Spilites

	1	2	3	4	5	6	7
Li		10	60	25	5		
Sc		10	18	18	18		
V	490	38	85	85	135	325	190
Cr	265	275	113	113	138	45	70
Ni	355	450	100	30	70	275	290
Co		45	25	36	36	<50	<50
Cu		54	63	250	>250		120
Zn	265					175	265
Ga		<2	5	7	5		
Rb	22	<5	50	<5	<5	35	30
Sr	90					60	250
Y		<5	7	10	15		
Zr			16	33	33		
Mo	-					-	-
Ba	20	<4	500	50	4	<20	<20
Pb							33

1. Montana Melaphyre (5055).

2. Picrite basalt (Scott, 1951a) King Island.

3, 4, 5. Augite spilites from King Island (from Scott, 1951a).

6. Core of pillow type 1 (32401) King Island.

7. Spilite, Lynch Creek (31658).

Analyses by X-ray spectrograph (M. Solomon), except for Cu and Pb by Aust. Mineral Development Labs.

Zr but are otherwise similar to those from basaltic rocks.

To summarise, the Tasmanian spilites almost certainly are derived from a basaltic magma but the various treatments of the chemical data reveal that differentiation of this basalt followed neither strictly tholeiitic, alkaline or calc-alkaline trends, but apparently some combination of all three. The problem will be further examined when other members of the Cambrian volcanics are considered.

ALBITE ANDESITES

The Crown Hill area, north of Queenstown, contains the only major development of andesites though they have been found north of Lake Dora (31793), and between Mt. Dundas and the Henty River bridge (62-86). Their total volume is estimated at about 1% of the volume of the spilites and considerably less than 1% of the volume of keratophyres. Their presence is of some importance to the problems of differentiation of spilites.

The andesites are characteristically porphyritic with phenocrysts of albite, hornblende, and quartz in a groundmass of albite, chlorite, calcite, epidote and iron oxides with rare apatite and sphene. In 31771A the iron ore is magnetite (X-ray diffraction).

The albite is present as laths up to 2½ mm. and generally much of it is very sericitised with only faint remains of twinning. However, some specimens show relatively fresh material in which the feldspar is loaded with inclusions of epidote (granular and "dirty"), chlorite and calcite. Zoning is shown up by alteration and in several cases a clear rim surrounds a sericitic core. The feldspar is generally low-temperature albite (An_{0-5}) but in one rock (32519) the feldspar was determined as andesine (An_{44}) by twin-axis measurements and extinction angles (Fig. 10). As well as the altered andesine laths very sericitised feldspar laths of unknown type occur in the same rock. This is the first definite report of calcic feldspar in the Cambrian lavas (but see p. 48). There is no sign of alteration to albite and the relationship of this type to the adjacent andesites is not clear because the rock occurs as an isolated boulder in the Lake Margaret moraine, 1½ miles north of Crown Hill. It may have been picked up locally or may have travelled across the West Coast Range from the Lake Dora area, where pebbles of hornblende andesite have been found.

The hornblende crystals are up to 2 cm. x 1 cm. in size and many show typical "primary" crystal outlines. However, in some thin sections the hornblende is much corroded, has a hazy brown fringe, and many inclusions of the groundmass and altered feldspar.

Scott (1952b) stated that there is no primary hornblende and that the crystals grow our of the alteration products of augite. However the three photomicrographs she figures do not necessarily demonstrate this at all. In 32524, fresh hornblende (euhedral and corroded) and heavily chloritised hornblende and chloritic-hematite laths all occur in one slide and probably represent the alteration of primary hornblendes. In other slides containing augite similar chlorite appears to be replacing augite. Yet this is not evidence that hornblende grows out of chlorite - rather the reve rse.

In 32519, a hornblende crystal appears to be derived by replacement of augite (see also 32865) and it seems likely that hornblende can be derived by "primary" crystallisation, or by crystallisation following or contemporaneous with resorption of augite. The embayed margins, inclusions of groundmass and albite laths, and "reaction rims" on many volcanic hornblende crystals are generally believed to be due to corrosion by residual liquid but may also be due to crystallisation from a crystal "mush", in which the later minerals might be expected to develop large inclusions and irregularities in shape.

Hornblende may be the only ferromagnesian mineral and up to 20% of the rock may be markedly subordinate to augite. This is present as cracked, stumpy crystals, partly altered to chlorite and similar to the diopsidic augite in the spilites (2 V measurements range from 52 to 61^{o}). As the percentage of ferromagnesian minerals falls, rocks tend towards the keratophyres with only a little chlorite and a few hornblende crystals (e.g. 31020). With decreasing hornblende and increasing augite and iron ores, they pass to basaltic rocks though a complete transition to spilite has not been observed. The hornblendes of different specimens (32522, 32519, 11912) give consistent optical properties, which are as follows:

Pleochroism:	X:	Pale straw-yellow
	Y:	Fairly pale yellowish green
	Z:	Medium grass-green
2V	:	73^{o}
$Z \wedge C$	:	18^{o} - 24^{o}
α	:	1.653
γ	:	1.676
$\gamma - \alpha$	:	0.023

This indicates substitution of about 40% Mg by Fe^{+2} + Fe^{+3} + Mn, according to Deer, Howie and Zussman (1962, 2, p. 296).

Quartz generally occupies 3-6% of the andesites, mainly in large cracked and corroded phenocrysts, also with chlorite in amygdules, and in irregular patches in the groundmass. Amygdules are less common than in spilites; they may be filled by chlorite, albite, quartz, epidote and calcite but are generally albite or an albite-chlorite-quartz combination. Albite, chlorite and quartz also occur in irregular masses in the rocks and probably formed during the later phases of lava solidification. Similar patches of pyrite and epidote granules are less common. Euhedral crystals of apatite are characteristic of the andesites and coarse crystals of magnetite occur sporadically.

The groundmass generally occupies 50-60% of the rock in porphyritic types, much less in seriate types. It is either a granular aggregate of albite or shows hyalopilitic texture, with albite microlites and dark interstitial material with weak birefringence that may be devitrified glass. Fine-grained chlorite, epidote and iron ores are intermixed with the albite and in one slide small hornblende laths occur. In specimen 62-86 myrmekitic intergrowth forms the groundmass of some of the rock; the texture is almost identical to that described by Kinkel et al. (1956, p. 26) in the Balaklala rhyolite. Albite laths show parallel orientation in some slides and this is probably associated with the parallel elongation of hornblende laths described by Scott (1954) and observed locally by the writer. This parallelism appears to be a flow phenomenon (Solomon, 1960) though Scott had some difficulty in accepting this because she believed that the hornblende crystals were post-solidification of the lava and formed by "regional metasomatism". The parallelism is not related to cleavage.

East of Crown Hill, the andesite is loaded with pebble-like fragments which consist of relatively fine-grained aggregates of hornblende and albite. They differ from the matrix in having finer grain, being richer in hornblende and having little or no groundmass. They probably represent crystal-clotting during solidification, probably in conjunction with some rotation of the "clots" that helps to compact them and to emphasise their outlines.

Considerable inhomogeneity is evident in the andesites, particularly with respect to albite content; several outcrops on Crown Hill show brecciation with angular blocks of andesite ~~and~~ in an andesitic matrix. Normally only visible on freshly spalled faces, this auto - brecciation (?) is highlighted in places by albitisation of the matrix, giving rise to spectacular outcrops (Plate 24, No. 1).

Numerous inclusions of andesitic lava and segregations and veins of epidote, albite-epidote and albite-epidote-quartz gives a very heterogeneous appearance to some outcrops. North of the Henty River inclusions consist of fine-grained, felted aggregates of albite laths with rare small, phenocrysts of albite and hornblende, suggestive of chilled material caught up in the main body of the flow.

Plate 24 No. 1:- Brecciated andesite, peak of Crown Hill.

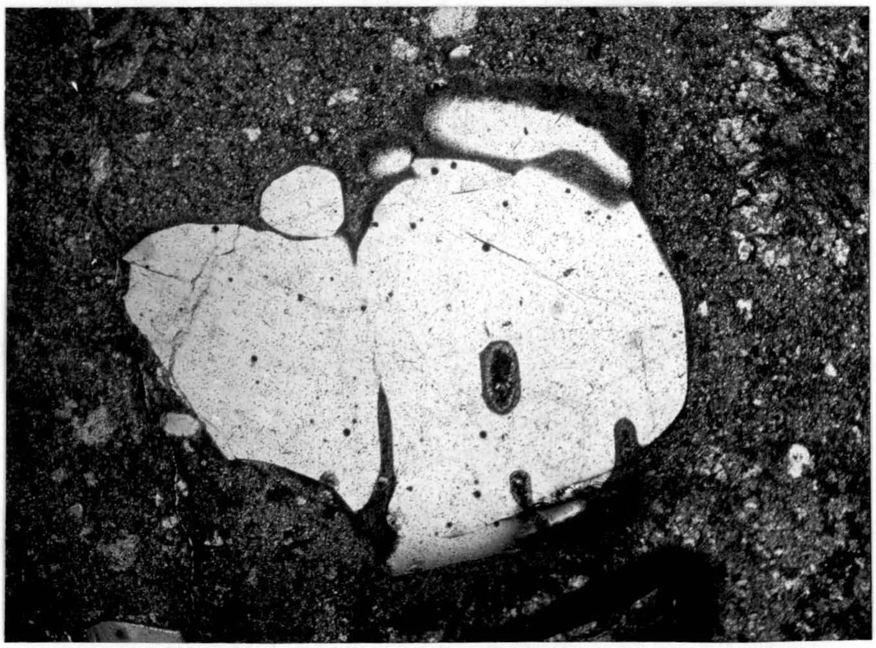


Plate 24 No. 2:- Embayed quartz phenocryst in quartz keratophyre, Red Hills. X35.

An analysis of hornblende-augite andesite with andesine and little albite (32519) is given in Table 7. The chemical analysis does not show P_2O_5 though crystals of apatite occur in several slides. The rock is unusually high in Sr, Ni and Cr.

Andesitic rocks are not common in spilite-keratophyre associations generally though Kinkel et al. (1956) described an albite andesite from Shasta Co., California, that is interbedded with spilite and keratophyres. They imply that the rock is an altered andesite (their "meta-andesite") but allow a possible primary origin for albite in related rocks. Bartrum (1929) gave an analysis of an "andesite" among spilitic rocks from New Zealand though it is obviously very similar to spilite and falls within the range of spilites as advocated by the writer.

Albite and albitised andesites are common in the Cascade volcanic province (Waters, 1955) in which andesites predominate over basalts and rhyolites. The more altered rocks are similar to those of Crown Hill in which parts are converted to albite and hornblende. These andesites are Eocene to Miocene in age and occupy several thousands of feet on the eastern margin of the Tertiary geosyncline in which were deposited also the Olympic Penninsula spilites and basalts of Eocene age. Waters believed the alteration followed

Table 7: Andesites

	1	2		3
SiO_2	61.12	61.58	V	170
TiO_2	0.42	0.49	Cr	170
Al_2O_3	17.65	16.96	Ni	265
Fe_2O_3	2.89	1.75	Cu	25
FeO	2.40	2.85	Zn	210, 245
MgO	2.44	3.67	Rb	40
CaO	5.80	6.28	Mo	-
Na_2O	3.83	3.94	Ba	1,122
K_2O	1.72	1.28	Pb	26
H_2O+	1.43	2.74		
H_2O-		nil		
MnO	0.15			
P_2O_5	0.15			
CO_2		0.25		
Total		100.35		

Table 7: Andesites (continued)

		2
qu		15.19
or		7.56
ab		33.30
an		24.85
di		3.81
hy	en	7.76
	fer	2.53
mag		2.54
calc		0.57
ilmen		0.93
		99.04

1. Average (52 analyses) hornblende andesite (Daly, 1933).
2. Hornblende andesite, Crown Hill, specimen 32519 (Solomon, 1960).
3. Trace elements in 32519 in p.p.m.

Trace elements by X-ray spectrograph (M. Solomon) except for Cu and Pb by Australian Mineral Development Labs.

burial of the lavas and depression of the geosyncline.

The volume of andesites in Tasmania is considerably less than (about 1%) that of the spilites and they might well be differentiates of basaltic magma. Judging by the position of the andesite on various differentiation curves (Figs. 15-19) the parent magma must be of calc-alkaline type.

KERATOPHYRES AND QUARTZ KERATOPHYRES

Nomenclature

The keratophyres consist mainly of albite with minor quartz, chlorite, magnetite, sericite, apatite etc. Examples from Mt. Read had been compared to the "original" keratophyres of Fichtelgebirge by Rosenbusch as early as 1899 (in Twelvetrees and Petterd). Rosenbusch pointed out that the German keratophyres had previously been known as porphyroids, a term which later was widely used in Tasmania (Scott, 1954; Bradley, 1954; Carey, 1953). The term keratophyre appears to have suffered less abuse than spilite, for the definition proposed by Wells (1922) appears to have been followed by the majority of workers. He stated that keratophyres were: "Intermediate members of the Spilitic Suite, both intrusive and extrusive, the rocks being characterised by a high percentage of sodic feldspar, usually anaacid plagioclase near albite in composition, accompanied by a small proportion of dark minerals, now represented by chloritic or serpentinous pseudomorphs." Wells implied elsewhere that the chlorite might also be primary. He added that the "quartz keratophyres are the corresponding acid rocks, carrying in addition notable amounts of free quartz". Battey (1955), following Beskow (1929) and Geijer (1931) and also Donnelly (1963), used the term quartz keratophyre only for rocks with phenocrystic quartz but the writer has preferred to use quartz keratophyre for the acid keratophyres with appreciable amounts of quartz, regardless of the presence or otherwise of quartz phenocrysts. The division is made at about 65% SiO_2 though this is obviously difficult to apply when the quartz is fine-grained and intimately mixed with feldspar, and no chemical analyses are available. A similar system has been adopted by Dickinson (1962) who says (p. 249) that quartz keratophyre is "composed dominantly of albite and quartz forming an interlocking groundmass mosaic.... Most quartz keratophyres that have been described may carry quartz and albite phenocrysts as well". Williams, Turner and Gilbert (1954, p. 101) suggest quartz is usually confined to the groundmass. These authors included phenocrysts of augite and hornblende in keratophyres and Gilluly

(1935, p. 228) mentioned hornblende. Von Gumbel's original description (1874) of keratophyres also mentioned occasional flakes of mica and hornblende.

Keratophyres are distinguished from sodic trachytes by the following criteria;

(a) Keratophyres do not contain the sodic amphiboles or pyroxenes characteristic of trachytes, and usually contain only chlorite.

(b) The plagioclase of keratophyres is invariably of the low-or near-low-temperature form but may be high temperature in trachytes.

(c) Feldspathoids and sanidine may occur in trachytes but not in keratophyres.

(d) Keratophyres are in many cases strongly metasomatised.

(e) Keratophyres tend to be associated with spilites or sodic granites, while trachytes accompany basalts and syenites.

Some confusion has appeared in earlier work involving sodic trachytes and keratophyres e.g. the rocks from Skomer (Wales) described by Thomas (1911) as soda-trachyte or "lime-bostonite" appear to be keratophyres. Thomas divided keratophyres from trachytes on the ground that the keratophyres have:

(a) less well developed flow structure,

(b) greater quantity of chlorite and sphene,

(c) better developed vesicular structure, and

(d) are "slightly more basic than the soda-trachytes and richer in lime and magnesia, ...".

As Wells (1922) points out, these factors probably only apply to the particular types studied by Thomas, and his chemical distinctions are based on comparisions with one analysis of a trachytic rock from North Pembrokeshire.

Johannsen (1937, p. 49) actually described sodic rhyolites and keratophyres together as a group and Harker, Tilley et al. (1960, p. 148) have described "common soda-rhyolites (sometimes called by the loosely defined name 'keratophyre' or 'quartz-keratophyre')".

In Daly's average keratophyre (1933, p. 12, 7 analyses), two possibly significant features stand out in comparison with his average alkaline trachyte - the lower Fe^{3+}/Fe^{2+} ratio and the higher water content. No change is indicated by Nockold's more recent (1954) average of alkaline trachyte. Combining these chemical features with the mineralogy, the Na-trachyte (analysis IV) of Thomas (1911) and the albitic porphyry of Bartrum (1936) and the lime-bostonite (analysis VI) of Thomas (1911) could be included among the keratophyres. Several keratophyre analyses (some of which were presumably included in Daly's average) are given in Table 8 (note that Battey's keratophyres are among the quartz keratophyres). The compilation of all pre-1961 Australian analyses of igneous rocks by Joplin (1963) has enabled a full coverage of Australian literature. One of Joplin's analyses of "keratophyres" (p. 139 no. 2) has been rejected - it is low in water content and comes from a granite contact aureole in New South Wales in which no other lavas are present. The average of all the analyses quoted is also included in Table 8. Though averages of this type suffer from possible misidentification by some workers, most of the examples in Table 8 have been checked against descriptions of mineralogy and field occurrence.

Unfortunately, a few of these keratophyres are rich in Fe_2O_3 and the hoped-for criterion of low Fe^{3+}/Fe^{2+} ratio becomes invalid. Similarly, a few of the keratophyres are low in water content. However, in general these two criteria are valid and help to identify the keratophyres in conjunction with the mineralogy. As Teall (1888) noted, the "original" keratophyres (which included quartz keratophyres) from the Fichtelgebirge (Gumbel, 1874) and the Harz Mts. (Lossen, in Teall, p. 370) have highly variable contents of potash and soda but for keratophyres in general, soda appears to dominate (see average in Table 8). If the keratophyres from the Middle Devonian of Lahnmulde (Gotz, 1937; Lehmann, 1949) are

Table 8: Keratophyres

	1	2	3	4	5	6
SiO_2	63.58	57.23	52.36	54.41	61.86	61.67
TiO_2	-	1.29	0.29	0.54	0.10	0.34
Al_2O_3	13.60	18.17	17.23	27.04	16.58	17.47
Fe_2O_3	6.71	1.02	4.13	1.88	2.17	1.37
FeO	4.47	4.96	7.53	0.11	6.39	3.92
MgO	2.58	1.47	3.18	0.51	0.54	2.13
CaO	-	1.19	4.29	2.00	0.28	0.18
Na_2O	5.25	4.67	5.10	1.14	4.98	8.52
K_2O	0.32	6.71	2.93	6.71	5.34	3.38
H_2O+ / H_2O-	2.94	3.00	3.01	3.36	1.95	0.45
MnO					Tr.	
P_2O_5		0.21	0.33	0.08	0.11	0.06
CO_2		0.01	0.21	1.22	0.30	0.05
FeS_2						
SO_3					0.12	
NiO		0.08	0.16			
BaO						
Ig. loss:						
Cl						
S				0.09		
ZrO_2						
SrO						
Total	99.45	100.01	101.77	100.44	100.72	99.59

Table 8: Keratophyres (continued)

	7	8	9	10	11	12
SiO_2	58.80	54.51	66.05	64.38	58.47	55.38
TiO_2	0.40	2.45	0.49		2.17	0.90
Al_2O_3	17.03	16.69	13.29	16.98	18.60	18.34
Fe_2O_3	2.44	2.49	3.22	4.04	1.92	1.13
FeO	5.81	5.31	5.07		4.77	5.86
MgO	1.83	0.67	1.36	0.28	0.94	3.47
CaO	1.16	2.01	0.50	1.08	0.99	3.25
Na_2O	5.22	1.36	6.67	7.57	5.52	7.12
K_2O	4.27	11.70	0.87	4.30	3.30	0.22
H_2O+	2.68	2.17	1.88		2.19	2.39
H_2O-			0.96		0.50	0.48
MnO					0.19	
P_2O_5	0.11	0.49	0.09		0.45	Tr.
CO_2	0.75				0.04	2.00
FeS_2						
SO_3	0.11	0.17				
NiO						
BaO					0.04	
Ig. loss				1.64		
Cl					0.02	
S						
ZrO_2						
SrO						
Total	101.47	100.21	100.45	100.27	100.11	100.54

Table 8: Keratophyres (continued)

	13	14	15	16	17	18
SiO_2	63.9	60.39	56.95	63.58	64.32	62.69
TiO_2	1.35	0.80	0.89	0.99		
Al_2O_3	11.9	18.40	17.87	13.42	18.46	14.83
Fe_2O_3	1.2	1.03	4.49	2.10	5.24	2.57
FeO	6.6	3.50	6.00	5.67	0.19	7.93
MgO	2.6	1.27	0.93	1.37	0.05	1.76
CaO	1.8	1.53	2.30	2.75	0.09	2.34
Na_2O	4.3	8.79	8.80	4.31	9.01	6.52
K_2O	0.25	0.46	0.38	2.93	0.32	0.63
H_2O+	2.2	1.37	0.71	1.85	1.82	0.73
H_2O-		0.20	0.38	0.30	0.27	–
MnO	0.39	0.08	0.08	0.14		
P_2O_5	0.64	0.12	Tr.	0.35		Tr.
CO_2		1.70	0.91	0.03	nil	
FeS_2		0.02	0.04			
SO_3						
NiO				0.01		
BaO				0.05		
Ig. loss						
Cl						Tr.
S				0.02		
ZrO_2				0.01		
SrO				0.01		
Total	97.13	99.66	100.73	99.82	99.77	100.00

Table 8: Keratophyres (continued)

	19	20	21	22	23	24
SiO_2	62.04	62.29	56.80	61.51	60.08 (21)	61.49
TiO_2	0.83	0.65	0.35	0.45	0.87 (17)	0.54
Al_2O_3	17.51	14.75	17.11	17.37	16.92 (21)	15.24
Fe_2O_3	3.97	2.29	2.12	1.92	2.74 (21)	3.44
FeO	0.74	3.02	5.26	3.35	4.66 (20)	3.73
MgO	1.43	1.99	3.61	1.26	1.62 (21)	0.95
CaO	2.68	4.77	4.20	1.08	1.97 (20)	0.72
Na_2O	7.05	1.75	4.47	5.23	5.62 (21)	4.99
K_2O	1.56	2.90	2.75	5.29	2.96 (21)	5.73
H_2O+	1.21		2.58	2.45	2.19 (19)	0.96
H_2O-	0.04		0.04			0.15
MnO		0.07	0.28	0.01	0.14 (9)	0.19 (11)
P_2O_5	0.08	0.11	0.35	0.08	0.20 (18)	0.08 (7)
CO_2	1.08		0.44		0.62 (14)	2.57 (10)
Ig. loss		5.55				
S		0.15				0.08 (10)
Total	100.31	100.29	100.36	100.00		

Table 8: Keratophyres (continued)

		20	21	22	23
qu		26.90	4.51	5.13	9.35
or		17.14	16.25	31.26	17.49
ab		14.81	37.82	44.25	47.55
an		22.95	15.77	4.84	4.55
cor		0.32	1.00	1.27	2.81
hy	en	4.96	8.99	3.14	4.04
	fer	2.71	7.85	3.84	5.12
mag		3.32	3.07	2.78	3.97
calc		0.00	1.00	0.00	1.41
ap		0.26	0.82	0.10	0.47
ilmen		1.23	0.67	0.86	1.65
		93.97	97.75	9756	98.41

Table 8: Keratophyres

1-3. Keratophyres from the Harz Mts. (2 and 3) and Fichtelgeburge (1), in Teall (1888, p. 371-372).

4 and 5. Keratophyres from Westfalia (4) and the Harz Mts. (5), in Johannsen (1937, 3, p. 51).

6-8. Keratophyre (6) and Kalkikeratophyres from the Harz Mts., in Rosenbusch (1910, p. 343).

9. Keratophyres from Cornwall, England, quoted in Dewey and Flett (1911, p. 209).

10. Keratophyre from Hamilton Hill, Scotland, quoted by Dewey and Flett (1911, p. 209).

11. Soda trachyte, Skomer, Wales, in Thomas (1911, p. 192).

12. Lime bostonite, North Pembrokeshire, as quoted in Thomas (1911, p. 192).

13. Intermediate lava, Cader Idris, Wales, in Davies (1959, p. 197).

14 and 15. Keratophyre and Magnetite Keratophyre (15) from New South Wales, in Benson (1915, p. 139). Calcite in (15) removed by hand.

16. Albitic porphyry from Great King Island, New Zealand, in Bartrum (1936, p. 517).

17. Keratophyre from Binduli, West Australia, quoted in Joplin, (1963, p. 140).

18. Porphyrite, Victoria, in Joplin (1963, p. 140).

Table 8: Keratophyres

19. Keratophyre from Glarner Freiberge, in Amstutz (1954).

20. Porphyrite, from Stirling Valley Road, near Rosebery, in Finucane (1932).

21. Keratophyre, near Lake Margaret, Queenstown, in Solomon (1960).

22. Average of 7 analyses of keratophyre, in Daly (1933, p. 12).

23. Average of analyses 1-21 inclusive - figures in brackets indicate number of analyses available.

24. Average of 12 analyses of Lahn keratophyres (in Gotz, 1937, p. 188-189).

included, then the average potash content would be much higher because these keratophyres are consistently high in potash (an average of 12 analyses = 5.73%). However, these rocks have a low water content ($< 1.5\%$) and do not appear to be typical keratophyres; with the associated weilburgites, they form an unusual suite about which there is much controversy. A high potash keratophyre from Lake Dora has been placed in the quartz keratophyres though its silica content is just under 65%; this has been done because most of the potassic keratophyres in Tasmania are acid rather than intermediate and it is convenient to place them together in one table.

In distinguishing quartz keratophyres and soda-rhyolites, Wells (1922) maintained that the keratophyres have more alumina and less iron and soda but a study of available analyses does not uphold this. The main basis of distinction seems to be field association, albite content, paucity of ferromagnesian minerals, except chlorite, and in particular, lack of sodic ferromagnesians (Hatch, Wells and Wells, 1956, p. 223). However, such factors are often ignored - for instance, Kinkel et al. (1956) have given descriptions of the Balaklala rhyolite that fit quartz keratophyre perfectly.

A number of quartz keratophyre ($> 65\%$ SiO_2) analyses have been collected, by no means all of those available, and the range of composition is clearly considerable (Table 9). The average is similar to a sodic rhyolite. It is possible that several of these rocks have been wrongly identified but it is doubtful whether these would alter the general conclusion.

There do not seem to be any textures characteristics of the keratophyric types.

In general, quartz keratophyres are more common than keratophyres and a similar situation exists in Tasmania.

Most workers have concluded that the Tasmanian keratophyric rocks are igneous (volcanic and intrusive) but Scott (1954) and Bradley (1954) suggested they were the products of the regional metasomatism of spilites and sediments. The writer suggests they are volcanic because (a) they have textures typical of volcanic rocks (this is particularly true for some of the Primrose Volcanics), (b) they have chemical and mineralogical compositions similar to those of well-established suites of volcanic rocks, (c) some of them contain quartz crystals that are likely to be of high temperature origin and (d) the field evidence used by Bradley (1954) to demonstrate gradational metasomatism has proved to be invalid (Solomon, 1960).

Table 9: Quartz Keratophyres

	1	2	3	4	5	6
SiO_2	70.97	66.20	70.57	69.48	76.81	75.10
TiO_2	0.25	0.06	0.06	0.14	0.19	0.22
Al_2O_3	13.84	17.76	15.39	11.99	11.70	12.84
Fe_2O_3	3.21	1.32	2.77	2.54	0.35	0.70
FeO	0.78		1.81	2.46	0.39	1.36
MgO	0.20	0.08	1.52	1.16	0.30	0.30
CaO	1.26	0.25	1.66	1.72	0.30	0.32
Na_2O	6.27	3.00	2.61	3.33	2.73	5.12
K_2O	1.57	10.54	2.21	4.01	5.45	2.39
H_2O+	0.74	0.68	1.12	1.56	0.96	0.95
H_2O-				0.32	0.51	0.27
MnO	0.12		0.05	0.20	Tr.	0.04
P_2O_5	0.08		0.34	0.08	0.08	0.04
CO_2	0.79	0.03	0.24	1.34	0.02	0.03
FeS_2				0.05		
NiO				0.03	0.01	
BaO					0.05	0.05
S			0.09		0.02	
ZrO_2					0.01	
SrO					0.01	
						Rest: 0.09
Total	100.09	100.00	100.44	100.41	99.93	99.82

Table 9: Quartz Keratophyres (continued)

	7	8	9	10	11	12
SiO_2	70.26	71.23	71.34	74.80	76.07	76.08
TiO_2	0.68	0.47	0.36	0.01	0.14	0.15
Al_2O_3	12.58	13.74	13.66	13.59	12.50	12.02
Fe_2O_3	3.30	0.45	3.28	1.05	0.75	1.05
FeO	2.00	1.18	1.48	0.29	0.89	1.09
MgO	0.36	0.27	0.34	0.24	0.40	0.50
CaO	1.26	0.58	1.26	0.25	0.13	0.18
Na_2O	4.77	0.53	5.74	1.42	4.37	1.84
K_2O	2.88	10.31	0.87	7.10	3.31	4.99
H_2O+	0.97	0.58	1.28	0.97	1.08	1.44
H_2O-	0.27	0.06	0.41	0.33	0.22	0.31
MnO	Tr.		Tr.			
P_2O_5	0.52	0.13	0.05	0.08	0.03	0.02
CO_2						
FeS_2						
NiO						
BaO						
S						
ZrO_2						
SrO						
Total	99.85	99.53	100.07	100.13	99.89	99.67

Table 9: Quartz Keratophyres (continued)

	13	14	15	16	17	18
SiO_2	71.52	72.31	75.04	81.33	65.52	72.83
TiO_2	0.28	0.40	0.10	0.25	0.47	0.35
Al_2O_3	11.76	12.76	13.39	9.21	14.90	12.77
Fe_2O_3	1.52	1.94	1.61	1.09	6.25	2.55
FeO	3.44	1.26	0.37	0.74	1.55	1.15
MgO	1.18	1.32	0.18	0.40	1.10	1.45
CaO	2.72	0.10	0.40	0.25	0.64	2.07
Na_2O	5.05	3.69	6.36	3.25	4.85	4.38
K_2O	0.26	3.82	0.83	1.66	1.97	0.80
H_2O+	1.25	1.48	1.07	1.12	2.03	1.35
H_2O-	0.14	0.26	0.24	0.15	0.39	0.12
MnO	0.04	0.08	0.05	0.05	0.05	0.04
P_2O_5	0.20	0.15	0.08	0.04	0.16	0.10
CO_2	0.38	Tr.	0.10	0.10	nil	0.18
FeS_2	0.12					
NiO						
BaO		0.09		Tr. ?		
S						
ZrO_2						
SrO						
Total	99.86	99.66	99.82	99.64	99.88	100.14

Table 9: Quartz Keratophyres (continued)

	19	20	21	22	23	24
SiO_2	70.45	77.01	73.60	68.75	76.47	76.95
TiO_2	0.36	0.18	0.27	0.27		0.10
Al_2O_3	14.47	12.04	13.08	16.75	13.90	12.59
Fe_2O_3	1.25	0.67	1.16	0.48	0.18	0.58
FeO	2.07	1.51	2.69	1.72	0.40	0.22
MgO	1.38	1.29	1.32	0.83	1.74	0.26
CaO	0.85	0.86	0.22	0.89	0.19	0.26
Na_2O	5.99	4.58	3.93	6.95	4.76	2.32
K_2O	1.70	0.59	1.65	0.80	0.65	5.76
H_2O+	1.13	1.19	1.73	1.52	1.34	0.76
H_2O-	0.06	0.13	0.13	0.84	0.29	0.00
MnO	0.05	0.03	0.05	0.04		
P_2O_5	0.10		0.07	0.16		0.14
CO_2	0.22		0.12			0.00
FeS_2						
NiO						
BaO				0.03		
S						
ZrO_2						
SrO				0.03		
Total	100.08	100.08	100.02	100.06	99.92	99.95

<u>Table 9: Quartz Keratophyres</u> (continued)

	25	26	27	28	29
SiO_2	69.33	70.60	71.30	69.48	70.57
TiO_2	0.32	0.58	0.51	0.14	0.06
Al_2O_3	15.66	11.50	13.53	11.99	15.39
Fe_2O_3	2.47	3.02	2.33	2.54	2.77
FeO	0.32	3.16	1.75	2.46	1.81
MgO	0.47	0.12	0.70	1.16	1.52
CaO	1.41	3.00	0.67	1.72	1.66
Na_2O	2.05	3.26	5.77	3.33	2.61
K_2O	5.33	2.23	3.02	4.01	2.21
H_2O+	2.03	1.50	0.56	1.88	1.12
H_2O-	0.07				
MnO		0.14	0.07	0.20	0.05
P_2O_5	0.08	0.10	0.03	0.08	0.34
CO_2	0.54	0.90		1.34	0.24
FeS_2				0.05	
NiO				0.03	
BaO					
S					0.09
ZrO_2					
SrO					
Total	100.12	100.11	100.24	100.41	100.44

Table 9: Quartz Keratophyres (continued)

	30	31	32	33	34	35
SiO_2	67.50	70.60	72.88	78.40	73.62	73.44
TiO_2	0.34	0.40	0.37	0.24	0.29	0.33
Al_2O_3	11.12	13.31	13.76	10.65	12.55	14.18
Fe_2O_3	0.87	1.54	0.89	1.10	1.53	1.46
FeO	1.31	2.36	3.44	1.48	1.95	0.55
MgO	1.76	1.75	0.94	0.83	0.93	0.43
CaO	4.73	1.68	0.16	0.34	0.48	Nil
Na_2O	5.18	4.44	2.86	4.01	3.35	0.16
K_2O	1.15	2.03	2.63	1.59	4.27	8.05
H_2O+	0.17	1.46	2.04	1.46	1.08	1.38
H_2O-	0.26	0.06	0.06	Nil	Nil	0.08
MnO	0.35	0.05	0.02	Tr.	Tr.	Tr.
P_2O_5	0.14	0.08	0.09	0.09	0.09	0.10
CO_2	5.50	0.73	0.31	n.dt.	n.dt.	0.38
FeS_2				0.06	0.05	
S	0.07	0.07	Tr.			Tr.
	100.45	100.56	100.45	99.69	100.09	100.54

Table 9: Quartz Keratophyres (continued)

		30	31	32	33	34	35
qu		29.50	31.88	42.78	46.64	34.80	41.20
or		6.80	12.00	15.57	9.40	25.23	47.80
ab		44.00	37.57	24.10	33.93	28.35	1.05
an			3.20		1.10	1.79	
cor		1.33	2.64	6.22	1.93	1.76	5.20
hy	en	4.40	4.36	2.80	2.07	2.32	1.10
	fer	1.72	2.49	3.96	1.41	1.84	
mag		1.16	2.23	1.39	1.60	2.22	0.92
hem							0.80
calc		8.10	1.66				
ap		0.34	0.19	0.34	0.21	0.21	
ilmen		0.61	0.76	0.76	0.46	0.55	0.60
siderite				0.81			0.93
Total		97.96	98.98	98.73	98.75	99.07	99.60

Table 9: Quartz Keratophyres (continued)

	36	37	38	39	40	41
SiO_2	70.42	72.66	63.95	72.50	66.88	75.45
TiO_2	0.36	0.39	0.48	0.36	0.42	0.17
Al_2O_3	12.31	14.60	16.57	14.42	11.17	13.11
Fe_2O_3	2.79	1.50	3.26	1.71	3.49	1.14
FeO	3.27	0.64	1.67	2.11	9.29	0.66
MgO	1.06	1.11	1.76	0.70	1.93	0.34
CaO	0.44	0.81	1.33	0.34	0.32	0.83
Na_2O	0.29	1.51	1.27	3.60	0.15	5.88
K_2O	7.69	4.47	6.11	2.20	2.07	1.26
H_2O+	1.43	0.99	1.98	1.99	3.30	0.69
H_2O-	Nil	0.74	0.49	0.19	0.44	
MnO	0.05	0.01	0.29	0.02	0.23	0.29
P_2O_5	0.11	0.11	0.19	0.07	0.09	0.18
CO_2	Tr.	0.34	0.46			
FeS_2	0.08					
S		0.04	0.08			
Total	100.30	99.95	99.98	100.21	99.78	100.00

Table 9: Quartz Keratophyres (continued)

	36	37	38	39	40	41
qu	35.63	44.62	29.44	40.70	48.44	34.41
or	45.44	24.42	36.11	13.00	12.23	7.45
ab	2.45	12.78	10.75	30.46	1.27	49.76
an	1.46	1.15	2.45	1.23	1.00	2.94
cor	2.97	6.86	6.97	5.67	8.32	1.00
hy en	2.64	2.77	4.38	1.74	4.81	0.85
hy fer	3.20		0.12	1.91	13.91	0.53
mag	4.05	0.97	4.73	2.48	5.06	1.65
hem		0.83				
calc		0.77	1.05			
ap	0.26	0.26	0.44	0.16	0.21	0.42
ilmen	0.68	0.74	0.91	0.68	0.80	0.32
Total	98.79	96.17	97.35	98.03	96.04	99.33

Table 9: Quartz Keratophyres (continued)

	42		43		44	
SiO_2	69.36		72.29		71.17	
TiO_2	0.68		0.28	(38)	0.36	
Al_2O_3	13.36		13.69		13.15	
Fe_2O_3	3.02		1.57	(42)	1.83	
FeO	1.70		1.42	(41)	2.55	
MgO	0.49		0.72	(42)	1.17	
CaO	0.40		1.05		1.00	
Na_2O	4.69		4.35		2.44	
K_2O	4.58		2.76		3.84	
H_2O+	0.65		1.25	(42)	1.78	
H_2O-	0.09					
MnO	0.07	(7)	0.08	(31)	0.09	
P_2O_5	0.09	(9)	0.12	(35)	0.15	
CO_2	0.10	(11)	0.41	(28)	0.97	(8)
S	0.18	(10)				

Table 9: Quartz Keratophyres (continued)

		42	43	44
qu		23.08	34.11	36.33
or		27.24	16.31	23.76
ab		40.35	36.81	22.59
an		0.56	1.83	4.59
cor		0.30	2.88	2.93
hy	en	1.20	1.79	2.72
	fer		1.00	1.64
mag		3.71	2.28	2.41
hem		0.48		
calc		0.20	0.93	
ap		0.34	0.28	0.26
ilmen		1.37	0.53	0.68

Table 9: Quartz Keratophyres

1. From the Harz Mts., in Teall(1899. p. 371-372).

2. From the Harz Mts., in Rosenbusch (1910, p. 343).

3 and 4. From Ecuador (3) and England (4), in Johannsen (1937, p.51).

5. From Mt. Camel, New Zealand (quoted in Battey, 1955, p.117).

6. From Great Island, New Zealand (Bartrum, 1936).

7-12. From North Island, New Zealand (Battey, 1955,p. 116-117).

13. From New South Wales (Benson, 1915, p. 602).

14-16. From Eastern Oregon (Gilluly, 1935, p. 235).

17. From Shasta County, California, the "Dekkas Andesite" of Albers et al. (1961, p. 27).

18 and 19. From Shasta County, California, the "Bully Hill rhyolite", of Albers et al. (1961, p. 32).

20-23. From Shasta County, California, the "Balaklala rhyolite" of Kinkel et al. (1956, p. 23).

24 and 25. From Glarner Freiberge, Switzerland, as in Amstutz (1954, p. 113).

26-29. From Mysore State, India, as in Pichamuthu (1946).

30. From the Great Lyell Shaft area (east of Queenstown); core from Mt. Lyell Co. borehole GL5 at 135 ft. (47 m.), specimen 31702. Analyst: Japan Analytical Chemistry Research Institute.

31. From West Queen River, lower dam site (Solomon, 1960, p.49).

32. From East Queen River, Comstock tram-line 800 m. north of lower zig-zag (Solomon, 1960, p. 49).

Table 9: Quartz Keratophyres (continued)

33. From Waterfall Gully, north of Great Lyell Shaft, specimen 31272. Analyst: Dept. of Mines, Tasmania.

34. From head of Whip Spur, specimen 31676. Analyst: Dept. of Mines, Tasmania.

35. From Intercolonial Spur, between Mt. Jukes and Mt. Darwin (Solomon, 1960, p. 49).

36. From Mt. Sedgwick, south-eastern slopes, specimen 31257. Analyst: Dept. of Mines, Tasmania.

37. From below Yolande River bridge on Zeehan-Queenstown road, specimen 31717A. Analyst: Japan Analytical Chemistry Research Institute.

38. From 200 m. west of Lake Dora, specimen 31707A. Analyst: Japan Analytical Chemistry Research Institute.

39. From Comstock tram line, just north of lower zig-zag, Mt. Lyell, specimen 31754. Analyst: Dept. of Mines, Tasmania.

40. Magnetite quartz keratophyre, Murchison River gorge, specimen 30047. Analyst: Dept. of Mines, Tasmania.

41. Average of 13 analyses as in Daly (1933, p. 10).

42. Average of 12 quartz keratophyres from the Lahn area, Germany (analyses from Gotz, 1937, p. 188-189).

43. Average of 45 analyses: 1-29 inclusive and also 16 analyses of Australian quartz keratophyres from Joplin (1963), p. 112, no. 79; p. 113, nos. 84 and 89; p. 114, nos. 93, 94 and 97; p. 121, nos. 46, 47, 48, 49; p. 122 nos. 56, 57, 60; p. 125, no. 19; p. 126, no. 33; p. 128, no. 18. Figures in brackets show number of analyses used where less than 45.

44. Average of Tasmanian analyses (30) - (41) inc. Figures in brackets show number of analyses used where less than 12.

Field Occurrence

The keratophyres and quartz keratophyres form massive outcrops that in many areas yield little evidence of the orientation, thickness and extent of individual bodies. They are well exposed along the Lyell Comstock road (close to the East Queen River) and rock boundaries can be determined by composition changes and the presence of thin beds of tuff and mudstone. The keratophyres of of this area appear to be conformable units (flows ?) up to about 100 m. thick. Bedded flows are also well exposed west of Mt. Tyndall and on the western slopes of Mt. Farrell north of the Murchison River.

In some areas e.g. south west of Mt. Sedgwick, in the Andrew River near Darwin and near the junction of the King River and the Queen River, there are very large bodies of quartz keratophyre with more or less circular outcrops. The field relationships are obscured by vegetation but it is possible that these are shallow intrusions (or vents ?).

A brecciated appearance on weathered faces is not uncommon in the quartz keratophyres, the angular and rounded fragments generally being of similar composition to the matrix. The random distribution and the variation in fragment size in many cases give the rocks a conglomeratic appearance. The origin of these breccias will be discussed separately.

Mineralogy of the Keratophyres

Most of the keratophyres are albite porphyries (or porphyrites) with variable but minor quantities of chlorite, quartz, magnetite, epidote, calcite and sphene. They vary considerably in texture but are generally porphyritic (e.g. 32529) with phenocrysts of albite up to 2 x 1 mm. in a fine-grained or microcrystalline groundmass. Some specimens show seriate texture (e.g. 32535) and some are composed almost entirely of randomly orientated albite laths in a crude bostonitic texture. Rocks transitional to hornblende and augite andesite contain a few phenocrysts of these minerals, generally considerably chloritised; in 32535, augite laths are clearly altering to hornblende and specimen 640, near Waratah, contains a few augite crystals up to 0.4 x 0.1 mm., and augite pseudomorphed by pleochroic chlorite. Ragged patches and slivers of chlorite may in some cases be relict amphibole or pyroxene but in most cases it appears to be a primary mineral. Glomeroporphyritic texture is common, with irregular patches or clots up to 1 cm. across composed of random albite laths and interstitial chlorite. Euhedral apatite crystals up to 1/3 mm. in length are characteristic of the keratophyres.

The groundmass varies from an almost equigranular mosaic of albite and possibly quartz with varying amounts of interstitial chlorite, sericite, iron oxides, epidote etc., to an extremely heterogeneous aggregate varying in grain-size and composition. Crudely outlined patches or clots of coarser albite that appear to have developed from the groundmass are characteristic of the latter type, and in many cases chlorite and magnetite also develop in these more coarsely grained patches.

In specimen 31660, the feldspars are completely clouded and the reddish dusty groundmass is characterised by tiny ragged patches of microcrystalline quartz aggregate - some show portion of a "cross" on rotating under crossed nicols.

In 32529, magnetite occurs as rare phenocrysts and locally in irregular masses; the rock could be called a magnetite keratophyre, similar to that described by Benson (1915) from New South Wales.

Albite occurs in three gnerations:

(a) as phenocrysts,

(b) in the groundmass, and

(c) in amygdules, clots and veinlets.

(a) The albite of the phenocrysts forms subhedral, stumpy laths, exhibiting albite, Carlsbad, pericline and Albite-Carlsbad twinning and showing varying types of alteration. Some are seriticised, others are replaced by granular epidote (615) and all have a "dusty" appearance - in some rocks the albite has a characteristic pinky-brown colouration as if it contained very finely divided hematite "dust". The albite is of low-temperature type and its composition is similar to that in the spilites, though twin-axis orientations indicate compositions varying between An_0 and An_{10}. However this wider range of composition may reflect only the larger number of determinations available for the keratophyres. The albite of the quartz keratophyres is similar to that in the keratophyres and the following remarks apply to all albite crystals in both the volcanics and related intrusives.

Determination of the composition by twin axis orientation measurements (using the curves of Slemmons, 1962) is hampered by alteration of, or number of inclusions in, the crystals and by their low birefringence. In addition, determination of the twin axis is in many cases difficult because of the low angles between great circles and the low angles between the X-axis of twinned individuals. As a result the plotted measurements, except for a very accurate 10% or so, are considered to be accurate to only $\pm 2^{\circ}$. Even so, the spread of points about the low-temperature curve is considerable (Fig. 9). Plotting of 2V measurements for analysed crystals also indicates somewhat anomalous optics and a histogram of all 2V measurements indicates a wide spread (Fig. 10). Some crystals

Fig. 9: Albite twin-axis orientations according to the method of Turner (1947) but using the new data of Slemmons (1962).

a: Carlsbad twins

b: Normal Albite twins

c: Pericline twins

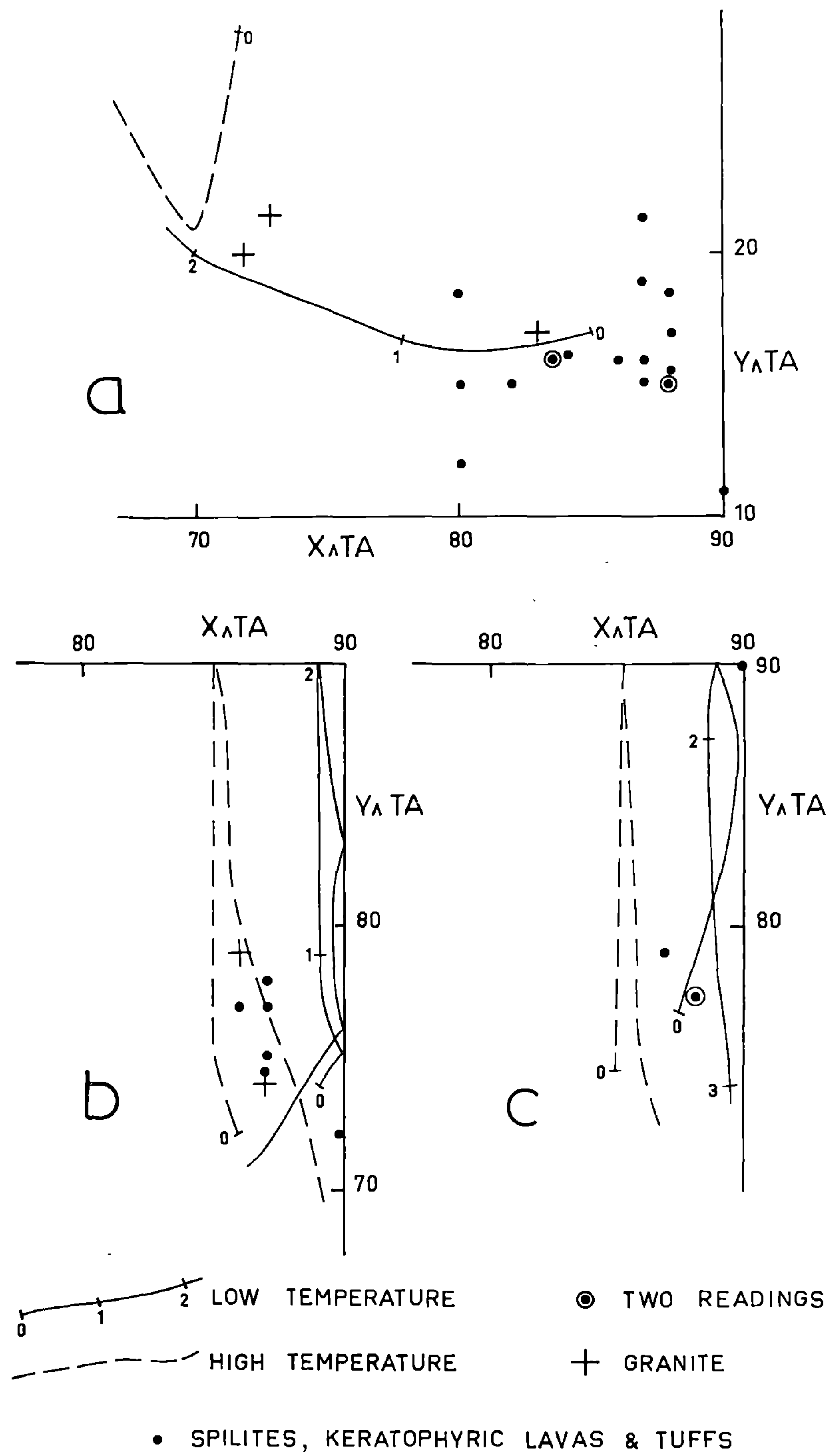
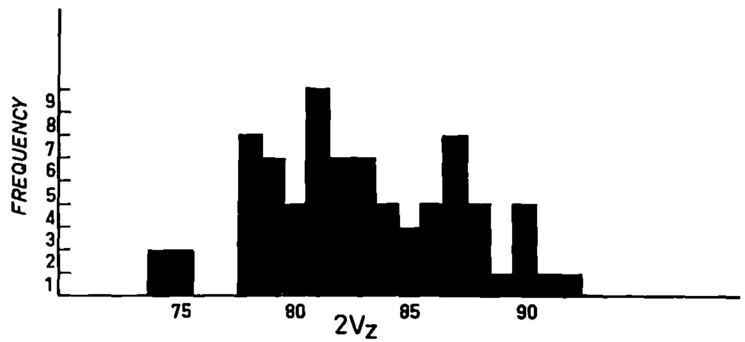
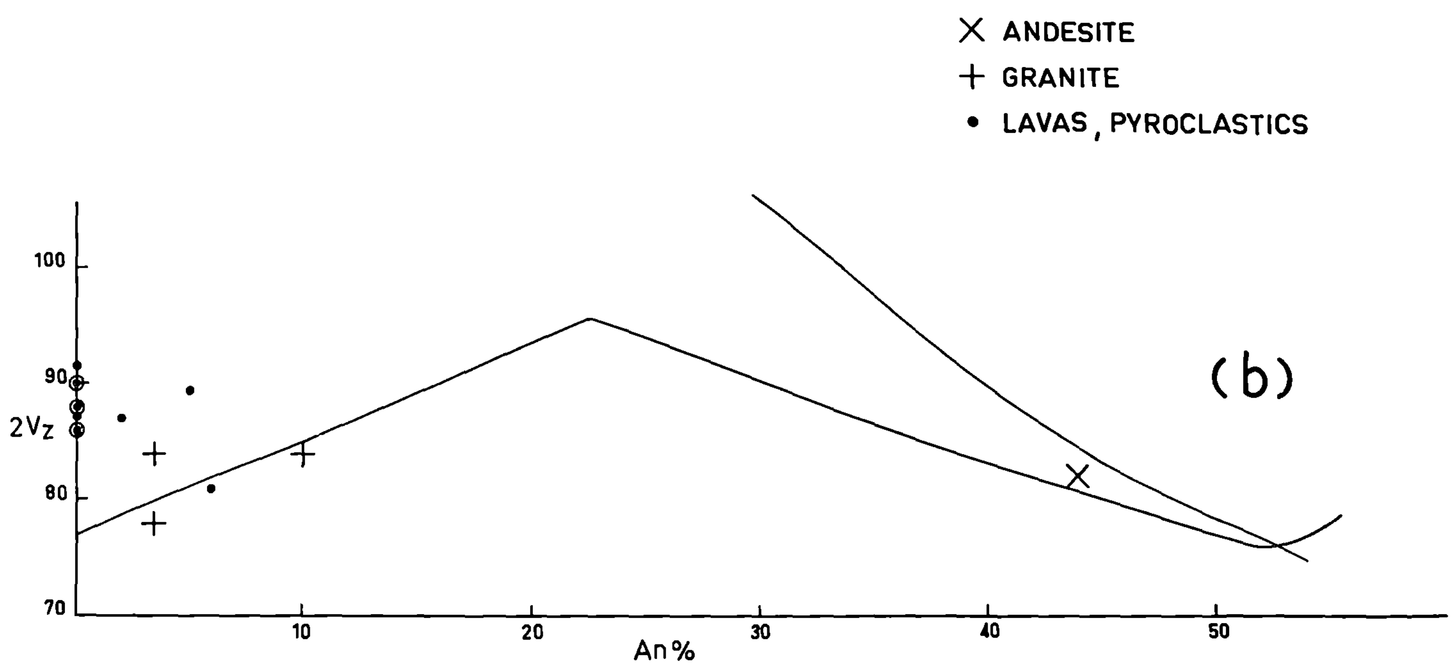

a
X∧TA
Y∧TA
70
80
90
10
20
b
X∧TA
Y∧TA
80
90
80
70
c
X∧TA
Y∧TA
80
90
90
80
LOW TEMPERATURE
HIGH TEMPERATURE
TWO READINGS
GRANITE
SPILITES, KERATOPHYRIC LAVAS & TUFFS

Fig. 10: Optical axial angle measurements (a) for all albites, (b) for albite in which composition is read from twin axis measurements; curves for high (upper) and low-temperature plagioclase taken from Smith (1956).



FREQUENCY
1
2
3
4
5
6
7
8
9
75
80
85
90
2Vz

(a)



ANDESITE
GRANITE
LAVAS, PYROCLASTICS
100
90
80
70
2Vz
10
20
30
40
50
An%

(b)

have what Donnelly (1963) describes as quasi-low temperature optics (QLT), with properties which lie between high and low temperature feldspars. The possible significance of this is discussed later.

Zoning is occasionally seen and some crystals have a rim which is generally clear albite. Inhomogenities in the albites are, however, rare and there is little or no sign of the replacement and exsolution textures that characterise the more acid rocks. The albite crystals are in general larger than those in spilites and in porphyritic varieties may be as large as 4 x 2 mm.

(b) As far as identification and examination of the fine-grained material is possible, the groundmass albite appears to be similar to that of the phenocrysts.

(c) The albite of the clots and amygdules is generally fresh and clear or faintly brownish. Twinning is less common than in the phenocrysts and when developed, indicates compositions similar to that of the phenocrysts. Veinlets of albite are fairly common and the albite is like that in the amygdules.

Many of the keratophyres are amygdaloidal, the principal fillings being albite and chlorite, with the latter usually occupying the core. The amygdules are in most cases well-defined in thin section but in some slides there appears to be a transition between clear-cut amygdules and the albitic 'clots' or patches just described.

It is suggested that both features are attributed to late stage volcanic activity, the 'clots' representing parts of the groundmass that have crystallised at a slower rate than the remainder due to the presence of irregular channels in which gas flow is concentrated and in which, perhaps, the temperature reamins abnormally elevated. In 31753, amygdules up to 1 mm. diameter contain calcite, quartz and chlorite and a few albite crystals; the carbonate occupies the core of the amygdule. In 31752, a lenticular vein (or amygdule ?) consists of euhedral quartz crystals and a randomly orientated sericite-chlorite aggregate which merges with the groundmass. This feature appears to be primary and indicates that much of the seriticisation of the rock, both groundmass and phenocrysts, may be of volcanic origin.

Chlorite crystallises later than albite (unless the blebs of chlorite in albite laths are early) and is interstitial to the feldspar. Primary chlorite crystallising later than feldspar has been noted in several keratophyres (e.g. Lehmann, 1949). It is generally pale to medium green in colour, feebly pleochroic and has very low birefringence. The later chlorite of the amygdules and vein-like segregations is similar in appearance. A typical segregation in a keratophyre from the East Queen River (31004) is composed entirely of a fine-grained aggregate of chlorite sheaves and plates, and the following properties were determined:

Pleochroism:	X	:	green
	Y	:	slightly paler green (?)
	Z	:	very pale greenish yellow
	2V	:	$0^{o} - 10^{o}$, - ve
		:	1.631
		:	1.633
		:	0.002
	S.G.	:	3.0

Sufficient material was available for analysis (Table 10) and the structural formula for a half cell, based on a unit cell of $O_{20}(OH)_{16}$, is :

$$(Al_{1.25}\ Fe^{+3}_{0.26}\ Fe^{+2}_{2.14}\ Mg_{2.23})\ (Si_{2.72}\ Al_{1.28})\ O_{10.29}\ (OH)_{7.71}$$

which classifies the chlorite as a ripidolite (Foster, 1962). The plot of this chlorite on the classification diagram used by Hey (1954) also falls in the ripidolite field and fairly close to the point indicated by the optical properties.

Chlorite also occurs in transitional keratophyres as replacements of augite and hornblende (e.g. 32536). Clusters of magnetite granules are in many rocks closely associated with chlorite slivers;

Table 10

SiO_2	25.82
TiO_2	Nil
Al_2O_3	20.46
Fe_2O_3	3.85
FeO	24.30
MgO	14.21
CaO	0.12
Na_2O	0.03
K_2O	0.02
H_2O+	10.88
H_2O-	0.61
MnO	0.15
P_2O_5	0.01
	100.46

Chlorite from keratophyre, East Queen River.

Analyst: Department of Mines, Tasmania.

the universal presence of magnetite with high-FeO chlorites in the keratophyres accounts for their low Fe^{3+} / Fe^{2+} ratios.

Mineralogy of the Quartz Keratophyres

In the field, the quartz keratophyres may not be readily distinguished from keratophyres unless phenocrystic quartz is present. The porphyritic varieties commonly have large, almost rounded, clear quartz phenocrysts that are as large as 1 cm. diameter (e.g. on Mt. Sedgwick and ½ mile south of Bulgobac); feldspar phenocrysts are invariably smaller. There are all gradations from coarsely porphyritic rocks to aphanitic types with quartz in a microcrystalline groundmass. The majority of the siliceous types are grey or brownish, massive rocks that form bold cliffs and tors.

The acidic group may be divided into a number of types that differ in field appearance, mineralogy and chemical composition, although there are gradations between types.

Sodic (albite) Quartz Keratophyres

Those with quartz phenocrysts are commonest and grade to non-porphyritic types. They consist of clear quartz and albite in a fine-grained quartz-feldspar groundmass. The quartz crystals are relatively clear and characteristically deeply embayed (Plate 24, No. 2). Some of them are well-formed bipyramidal crystals, and etching of basal polished sections of these crystals reveals pronounced shattering. These are features supposed to be typical of β -quartz that has cooled through the inversion temperature (Wright and Larsen, 1909) though Frondel (1945) has pointed out that such criteria cannot be rigorously applied. Polygonal cracking of the quartz crystals shows up in many thin sections and may be due to inversion to α -quartz during cooling; in some cases, groundmass material has penetrated the cracks.

The margins of the quartz crystals may be sharp and clear (32532) or have a blurred or "fussy" rim (32564, 639, 10, 3522 - Plate 24, No. 2). In general, where the "fussy" margin or rim is very thin, it appears to be highly sericitic. Where the rim is thicker the sericite flakes are perpendicular to the rim margin and fewer in number. Where the rim is about 0.02 mm. thick, it consists mainly of quartz (in optical continuity with the core crystal), with a few sericite (?) flecks and a seritic film at the crystal margin. Rim growth reaches a maximum in 32495 (Mt. Jukes), in which there has clearly been considerable overgrowth on the original euhedral crystals (Plate 25, No. 1); the overgrowth areas are loaded with

Plate 25 No. 1:- Overgrowth of quartz on primary crystal, Mt. Jukes. X65.

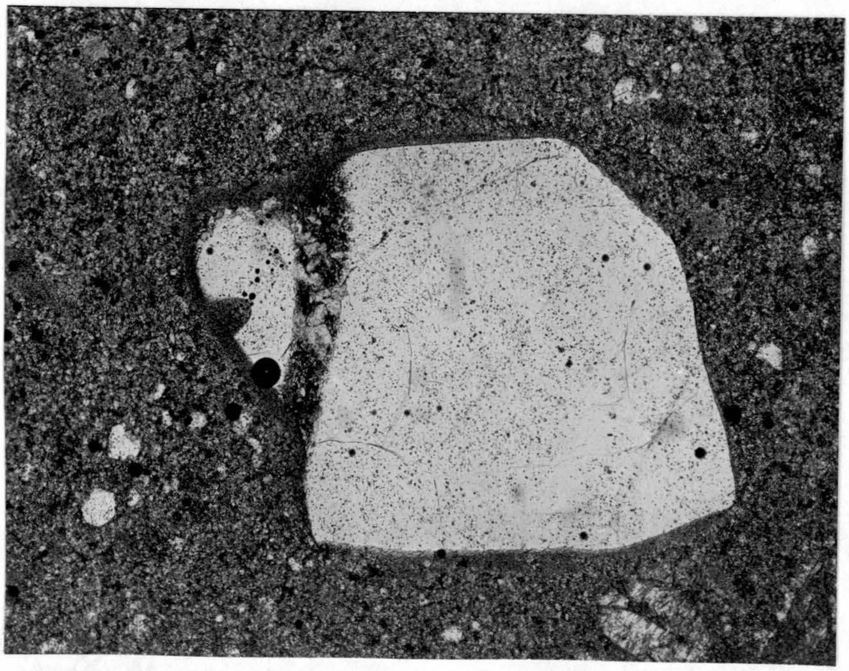


Plate 25 No. 2:- Rim surrounding veined quartz phenocryst, Red Hills. X35.

ragged sericitic (?) inclusions, have indecisive, fretted external margins, and are in optical continuity with the core. The crystals and overgrowths are considerably fractured and strained and are wrapped around by the later Devonian cleavage. The rims are clearly accretionary phenomena, growing on well developed crystal faces. They presumably formed by the gathering of silica from the groundmass and the removal of material other than silica, except for the initial sericite that developed at the primary crystal margins. The rims appear to follow around the margins of the "corrosion" embayments mentioned below and are presumably a late-stage phenomenon. The presence of sericite (?) within the rims suggests that some hydration of the groundmass feldspar may have occurred prior to rim growth. In specimen 3522 several quartz crystals have been fractured and moved prior to the rim growth and in one, the rim encloses two parts of a crystal that are separated by a chlorite-calcite vein (Plate 25, No. 2). Quartz rims upon feldspar crystals have been noted in only a few cases (e.g. 3510).

The presence of embayed margins or "corrosion" in quartz phenocrysts is a feature of many acid volcanic rocks. Foster (1960) has discussed the difficulties of understanding this and has rejected Moorhouse's (1959) suggestion that the crystals are "foreign" to the magma. Foster concluded that the crystals form early and react with a later liquid phase or with a vitreous groundmass. He suggested that the reaction between crystals and groundmass "probably takes place after the rock has solidified but while it is still hot". In 32525 it is fairly clear that some movement along cracks has occurred after embayment development. This "jostling" effect implies some fluidity in the groundmass and indicates that the rock was not solidified before corrosion. The penetration of groundmass into the cracks supports this inference. There is no way of telling whether the cracking preceded corrosion or not or whether the cracks are definitely due to $\beta - \alpha$ inversions or not. If they are shrinkage cracks due to inversion then some fluidity below about 573°C is implied.

Hydrous basaltic material might be expected to have a viscosity of say, 10^{10} at 500°C and more acid material an even higher viscosity. However, hydrous acid magmas almost certainly have lower viscosities for given temperatures though no figures are available to the writer.

In summary, rim growth followed corrosion, veining and some cracking and disturbance of crystals, but its time relationship to $\beta - \alpha$ inversion is not clear. There is some evidence that the lava remained partly fluid to a relatively low temperature and that corrosion embayments developed while the groundmass was still fluid.

Clearly there are at least three generations of quartz in the quartz keratophyres: as phenocrysts, in the groundmass and in amygdules and veinlets. The rim growth may be an additional phase.

A similar multiple origin for albite is seen in both keratophyres and quartz keratophyres. The albite, as in the keratophyres, is the low-temperature form and it occurs similarly in three generations. The feldspars of each generation have similar optical properties but that in amygdules and veinlets is invariably fresh, almost water clear and generally red in colour in contrast to the pinkish or greyish colour of the earlier phases. A few Manebach twins were identified in addition to the more common twins of the Carlsbad, albite, pericline and albite-Carlsbad laws.

Ferromagnesian minerals are scarce in the quartz keratophyres and are only found in quantity in transitional rocks with few quartz phenocrysts (e.g. 32538, 32532, 30050, 30051, 640). In some of these are found up to 10 or 15% of chlorite laths, probably pseudomorphing hornblende, and elongate laths that may have been biotite crystals.

Very variable amounts of late-stage calcite and sericite are present, but at least some calcite and sericite may be related to "burial" or later metamorphism. Euhedral apatite crystals are rare but characteristic. Epidote and epidote-quartz replacement is seen in places e.g. on the Murchison Highway, 1½ miles north of Tullah epidote blebs up to 10 cm. across occur scattered through quartz keratophyre (Plate 26, No. 1). Epidote is fairly common in veinlets with quartz. Though amygdules are rare in quartz keratophyres, they are prolific in some of the transitional types. In 32532 and 32926, 3 miles north of Queenstown, amygdules as large as 3 cm. diameter contain reddish albite, chlorite, quartz, calcite, epidote, and some have cores of galena and sphalerite (Plate 26, No. 2). This rock in bulk yields remarkably high values of lead and zinc (Table 13).

The groundmass, which forms the bulk of these rocks, is generally a confused aggregate of quartz, feldspar, chlorite, iron oxides etc., varying from microcrystalline to fine-grained.

Analyses of sodic quartz-keratophyres from near Queenstown and Rosebery are given in Table 9.

Potassic Quartz Keratophyres

Darwin Keratophyre

Spherical bodies of potassic (?) feldspar up to 0.5 mm. diameter make up a very important part of a particular keratophyre that crops out along the West Coast Range south of Queenstown. This

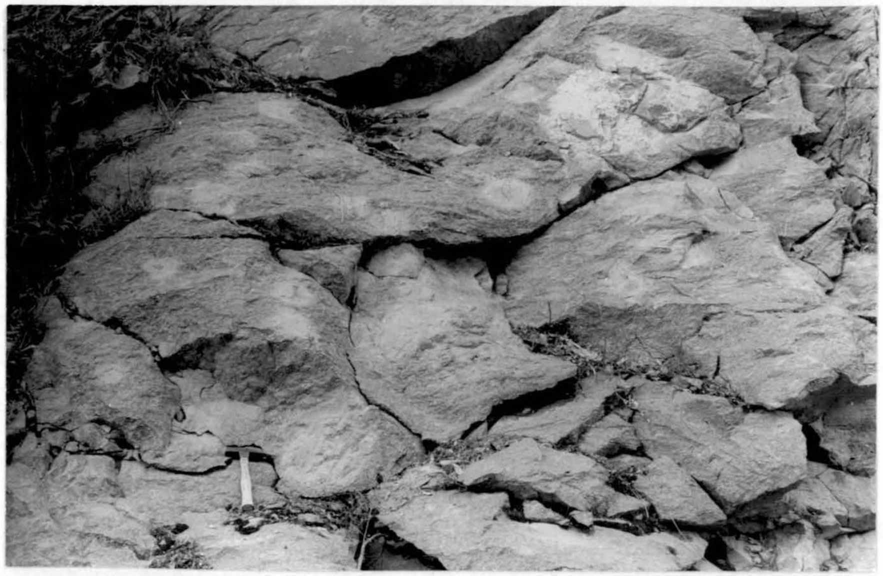


Plate 26 No. 1:- Epidote blebs in quartz keratophyre, north of Tullah.

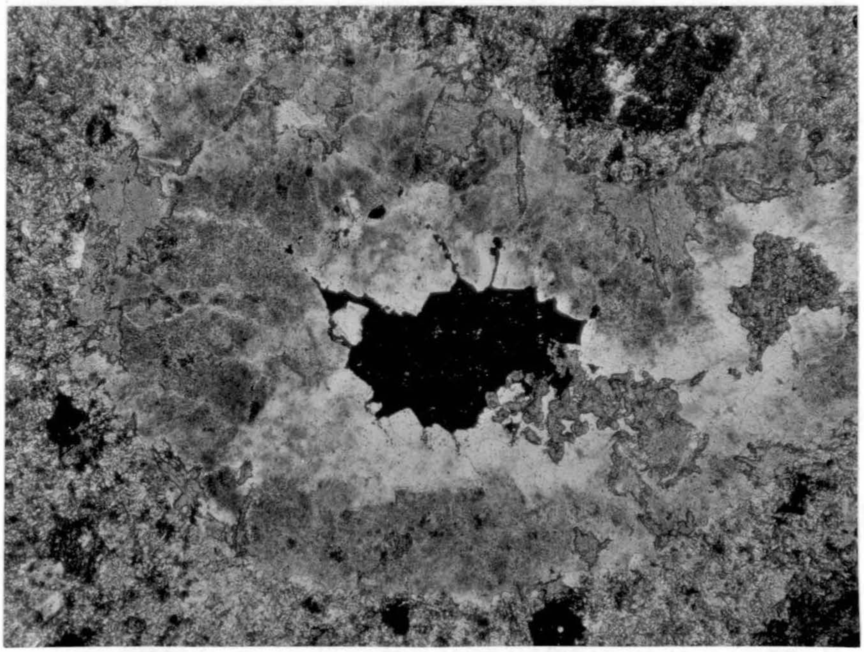
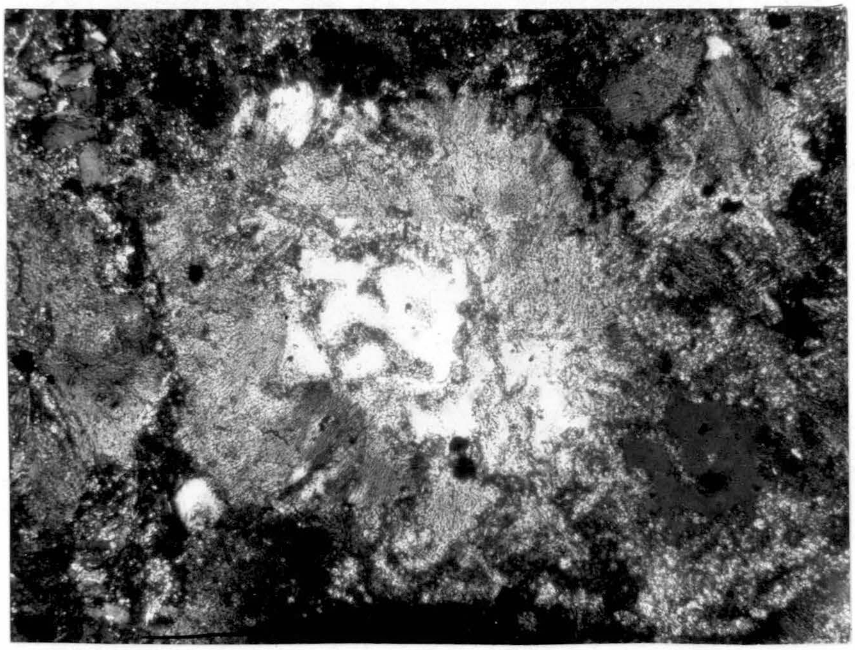


Plate 26 No. 2:- Sphalerite in core of amygdule, Zeehan-Queenstown road. X35.

rock, briefly described and termed a potash rhyolite in a previous thesis (Solomon, 1957) is distinctive in the field and forms resistant outcrops. For convenience, it is now termed the Darwin keratophyre. It is pinkish in colour, is closely and irregularly jointed and most of it is featureless. Parts of it are porphyritic but phenocrysts are rare, consisting mainly of albite and a few embayed quartz grains. Most of the rock (31769A, 32869, 32863, 32503) consists of radial feldspar masses centred on quartz grains and in places rimmed by quartz (Plate 27, Nos. 1 and 2). Seen at 1800 x enlargement the radial material appears to consist of almost vermiform stringers of crystalline material, with a crude radial alignment, and interstitial dark, very weakly birefringent material. The whole of the "spherulite" extinguishes together indicating that the masses are not truly spherulitic in structure. Other parts of the groundmass consist of irregular aggregates of feldspar (both albite and potassic) and quartz and sericitic material, with only rare radial growths. Granophyric intergrowth has been described by the writer (1957) in similar material from Lake Jukes (32862) and also occurs in rocks 400 m. north-east of Hercules (31788A).

Darwin keratophyre occurs at Mt. Sedgwick, Whip Spur, Red Hills (named after the reddish colour), east of Williamsford



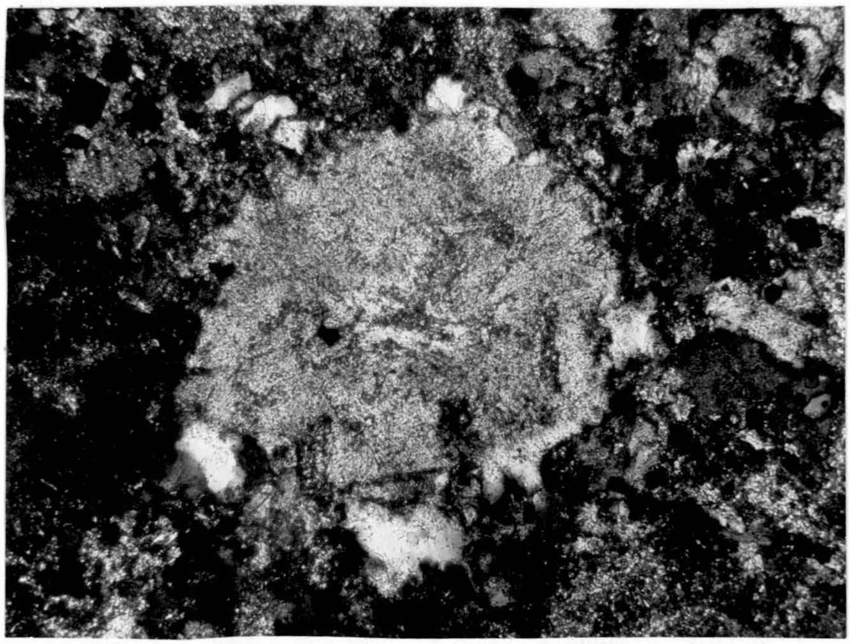


Plate 27 No. 1 and 2:- K-feldspar and quartz spherules in Darwin keratophyre. X135.

(31789A), Rosebery (32906), north of Lake Dora, 1 mile south of Tullah (31786), near the Pieman River railway bridge (32927), and in the Murchison River gorge.

Potassic keratophyre forms a plug-like mass south-east of Mt. Sedgwick (31257, 31253, 31254). It is identical in the field with the Darwin rock but it differs in having a few rather poorly developed crystals of K-feldspar phenocrysts as well as albite and quartz. A bulk analysis of this material, including variations in textural type, is given in Table 9, No. 36. The similar occurrence at Whip Spur (32522, 31676) which forms a bold outcrop overlooking Queenstown, differs in having no K-feldspar and more albite phenocrysts. This difference is reflected in the analysis (Table 9, No. 34). At Red Hills (3517, 3522) Darwin keratophyre forms a more or less concordant mass interbedded with banded siltstones and keratophyric rocks. Some of the rock has phenocrysts of K-feldspar in crystals that are up to 0.3 mm. diameter; they are characteristically crudely developed, seldom showing clear crystal faces, and many of them have a fringe of chlorite and magnetite. Most of them are heterogeneous and show wavy extinction, some are composed of sub-parallel crystals of slightly different optical orientation (Plate 28, No. 1), and a few show crude perthitic textures with sub-parallel stringers, blebs and laths (averaging 0.02 mm. diameter) of plagioclase feldspar (albite ?)

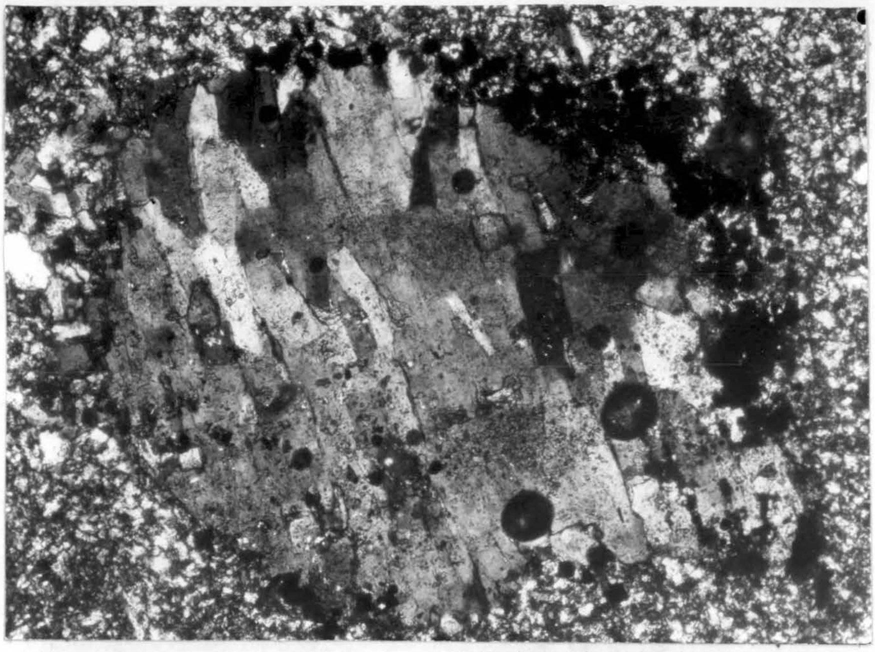


Plate 28 No. 1:- Heterogeneous K-feldspar phenocryst in Darwin keratophyre, Red Hills. X55

Plate 28 No. 2:- Veins of hematite and magnetite in Darwin keratophyre, Mt. Sedgwick.

in the K-feldspar host. The plagioclase bodies are approximately parallel to (100).

Irregular patches up to 1 mm. diameter consist of an aggregate of laths of K-feldspar with interstitial chlorite. The feldspar is relatively pure, homogeneous and in relatively well-formed laths compared to the scattered phenocrysts just described. It has a characteristic patchy, "turbid" appearance. The aggregates are probably recrystallised portions of the groundmass and appear to be of similar type to the phenocrysts.

Some Darwin-like keratophyres have a groundmass showing a "pellet-like" texture under low magnification (e.g. 32564, 3504, 3527). This is produced by the presence of numerous, crudely spherical bodies of potassic feldspar up to 0.03 mm. diameter which abut against each other to give the "pellet" appearance. Many of these bodies have radial texture and quartz cores, thus simulating the larger bodies in the normal Darwin keratophyre. In other Darwin types (e.g. 30051, 30059, 592, 1517) quartz and feldspar form a closely interlocking granular aggregate.

The Darwin keratophyres forms bodies of two characteristic shapes: tabular or dyke-like, and plug-like. All the bodies are tabular except those of Sedgwick and Whip Spur. The tabular bodies are probably flows (or very shallow intrusives) and the plug-like masses may be shallow vents or volcanic spines.

Determination of the composition of K-feldspars by optical studies is difficult if not impossible according to Hewlett (1959), but as White (1962) and Deer et al. (1962) point out, K-feldspar can at least be classified into one of the groups indicated by Tuttle (1952). Refractive indices give an approximate composition and 2V figures show the compositional "field". Accurate refractive index determinations proved extremely difficult because of the lack of clear crystals and because some variation between crystals appeared to be present. α lies between 1.518 and 1.520, similar to the K-feldspars in the Murchison and Darwin Granites, indicating a composition approaching 80-100% Or.

Measurements of the $(\bar{2}01)$ spacings (Bowen and Tuttle, 1950) indicated high-potash compositions.

2V figures for ten K-feldspars varied from 51° to 80°, the majority falling between 64 and 73°. The high and low values were recorded from the Sedgwick keratophyre in which the K-feldspar crystals are poorly developed and accurate readings for 2V are difficult. The 2V figures and refractive indices classify the feldspars as orthoclase and microcline cryptoperthites. X-ray diffractograms for Darwin keratophyres 31257, 32503 and 3517 (i.e. both with and without K-feldspar phenocrysts) are very characteristic and show the most intense quartz and orthoclase peaks but no albite peaks. This highlights the lack, or at best small amount, of perthite.

The normative values for the Sedgwick feldspars indicates the almost complete lack of albite and the presence of K-feldspar close to orthoclase.

The Tasmanian feldspars show some of the features regarded as typical of K-feldspars in volcanic rocks. Folk (1955) has found the clarity of the crystals a characteristic and White (1962) has added lack of twinning and association with high temperature plagioclase. Kohler (1949) suggested volcanic-type feldspars are euhedral and elongated along the a crystallographic axis. The Tasmanian crystals tend to be fairly clear (relative to the plagioclase feldspars) and they lack twinning but they are poorly develbped, show no particular elongation and are associated with low-temperature albite.

The only ferromagnesian mineral of the Darwin keratophyre is green, slightly pleochroic chlorite, generally present as small clusters of granular crystals, as radiating sheaves, or irregular masses. Pseudomorphs were not observed.

Apart from albite, orthoclase, quartz and chlorite, the only other characteristic minerals are euhedral apatite and magnetite and some hematite. The iron oxides occur as scattered crystals, as irregular masses in the groundmass and in veins. A hematitic K-feldspar porphyry (3523) from Red Hills consists of K-feldspars laths up to 0.5 x 0.3 mm. and ragged pathes of hematite up to 4 x 1.0 mm. in a microcrystalline quartzose aggregate. Specimens show faint parallel bands (1-2 mm. thick) of slightly different pinkish grey colours and the microscopic clots of hematite are generally streaked out parallel to the banding, which is probably a flow structure.

Magnetite-rich Darwin keratophyre occurs in the Murchison Gorge (30047) and consists of quartz and rarer K-feldspar phenocrysts in a devitrified, chloritic groundmass. Magnetite occurs in streaks and anhedral crystals, clearly pre-cleavage and pre-quartz veining. The analysis of this rock confirms the presence of magnetite and K-feldspar (Table 9,No. 40). Similar rock (30059) has a holocrystalline groundmass consisting of a quartz-K-feldspar aggregate.

Iron oxide veins up to 1 m. thick occur on Mt. Sedgwick where they contain chlorite, quartz, K-feldspar and fragments of keratophyre, making a fairly composite structure; fine veinlets also ramify the adjacent rocks (Plate 28, No. 2). The veins trend mainly about NNE though there is a considerable spread of orientations. Polished sections (Plate 29, No. 1) and X-ray diffractograms reveal that the

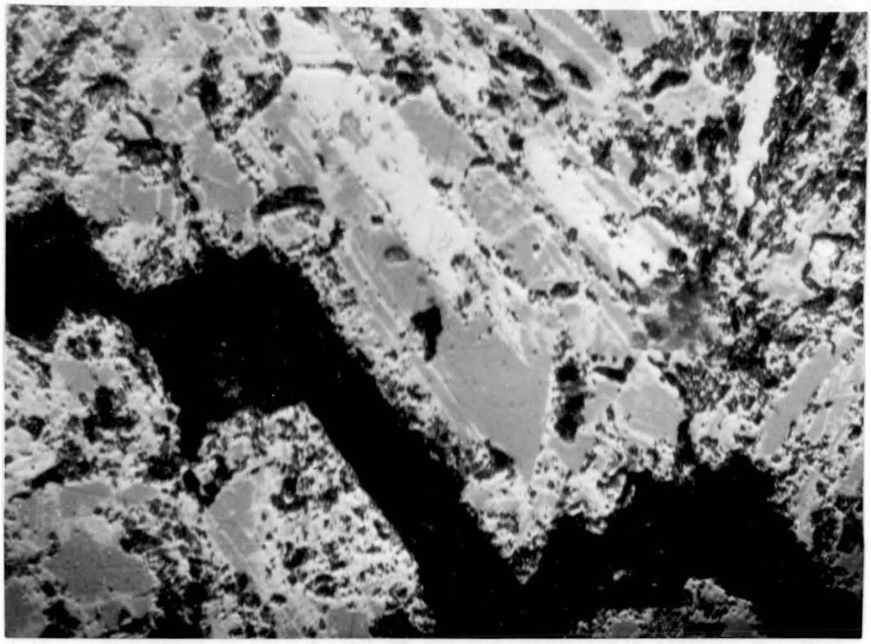


Plate 29 No. 1:- Magnetite partly oxidised to hematite in iron oxide veins in Darwin keratophyre, Mt. Sedgwick. X150.

Plate 29 No. 2:- Flow (?) banding in quartz keratophyre (?), East Queen River. Width of outcrop approximately 3 m.

iron oxides are largely hematite that show relics of an earlier magnetite; judging by the variation in magnetic intensity, the distribution of magnetite is erratic. Parts of the veins consist of specularite and these parts appear to be recrystallised.

Copper mineralization is associated with prominent hematite-magnetite veins between Mt. Darwin and South Darwin and also at Red Hills. On Mt. Darwin, several small copper deposits have been opened up and proved to be uneconomic. The largest is the Prince Darwin deposit (Fig. 3) which consists of a N-S stystem of veins of hematite and magnetite up to 70 m. wide. Underground workings show the iron ores to be speckled with pyrite and chalcopyrtie carrying silver and gold (Hills, 1914a). At Red Hills, several composite, N-trending hematite-magnetite veins, up to 50 m. thick, cut Darwin keratophyre. Shallow trenches and adits reveal pyrtie and chalcopyrite within the veins, in identical fashion to the Prince Darwin example.

Other Potassic Quartz Keratophyres

Potassic volcanics extend in a narrow belt from north of Rosebery to White Spur (south of the Hercules mine) and similar rocks have been found on and near Madam Howards Plains. Though some of these rocks may be lavas, many have fragmental texture and are probably tuffs and agglomerates. They are described in the sections on the fragmental volcanics.

Potassic lavas occur at Lake Dora (31707A) and in the Yolande River north of Madam Howard Plains (31717A). In the Yolande River, at the Queenstown road bridge, occur pale buff-grey massive quartz keratophyres in flows several tens of metres thick and interbedded with mudstones and fine-grained feldspathic tuffs. The rock is porphyritic, with subhedral quartz crystals up to 3 mm. across and a few smaller albite (?) crystals in a fine-grained aggregate (average 0.05 mm. diameter) of quartz, K-feldspar and sericite (?). The groundmass is vaguely similar to the "pellet" texture of the Darwin keratophyre and its composition is not dissimilar to some of the Darwin rocks (Table 9, No. 37).

West of Lake Dora there occur several hundred metres of featureless keratophyre, with some volcanic breccias and tuffs, underlying the Owen Conglomerate (Fig. 22). This keratophyre is similar to that from the Yolande River but its percentage of quartz phenocrysts varies considerably. An analysis (Table 9, No. 38) of a type almost free of these is actually a keratophyre ($< 65\%$ SiO_2) but most of the rock would be closer to a quartz keratophyre. It has a higher percentage of albite laths (about 10 %) than the Yolande rock and they are up to 1 mm. long. There are also several "laths" of chlorite pseudomorphing an earlier mineral and crystals of magnetite up to 0.3 mm. across. The groundmass is particularly "dusty" and confused but consists mainly of K-feldspar, some quartz and iron oxides.

The Lake Dora rock is the host to several small ore occurrences consisting of pyrite, chalcopyrite-pyrite or chalcopyrite-sphalerite-galena assemblages.

Sodi-Potassic Quartz Keratophyre (?)

A rock exposed in the upper reaches of the East Queen River has been previously described as a sodi-potassic rhyolite (Solomon, 1960). It is pale grey and in places is finely banded (Plate 29, No. 2), the banding in part being contorted (non-tectonic). This banding is invisible in thin section and may be due to flow. Microscopically, the rock is porphyritic with rare phenocrysts of albite in a fine-grained sericitic (?), quartz-feldspar (?) groundmass. The latter has a somewhat equigranular, detrital appearance and it is possible that this rock is a fine-grained tuff.

Environment of Deposition of the Keratophyric Lavas

Most of the lavas just described are from the Mt. Read Volcanic Arc and, in themselves, they yield little evidence of their environment. However, a study of interbedded fragmental rocks and sediments throws more light on this problem. For example, at Lake Dora lavas are associated with fine-grained sandstones showing convolute folding (Plate 44, Nos. 1 & 2) and clearly of aqueous origin. On the other hand, some of the pyroclastic rocks of the Primrose Volcanics are probably sub-aerial. It appears that the Mt. Read Volcanics are partly sub-aerial and partly sub-aqueous in origin.

The only keratophyric lavas found in the marine (?) sedimentary trough in which the spilites developed have been found about 4½ miles west of Waratah (639, 640) where they occur in a spilite-greywacke succession.

FRAGMENTAL VOLCANIC ROCKS

In describing these, the classification suggested by Fisher (1961), with modifications by Wright and Bowes (1963) and by the writer, has been used. It is shown in tabular form below:

Grain size (mm.)	Autoclastic			Pyroclastic
	Volcanic		Sub-volcanic	
	Flow-friction-explosion-breccia	Ignimbrites, froth flows	Flow-friction-explosion-intrusion-breccia	Breccia agglomerate
64				
				Lapillistone
2				
	Tuffisite			Coarse tuff
1/16				
				Fine tuff

A few words of explanation are necessary. The "epiclastic" group of Fisher is not included on the grounds that Wright and Bowes have rightly criticised the use of 'volcanic' as an adjective defining sandstone, conglomerate etc. (see also Williams, Turner and Gilbert, 1954), because a process is implied. Basaltic sandstone, rhyolitic sandstone, etc., are the correct terms for sedimentary rocks composed of the products of weathering and erosion of volcanic rocks.

The term "alloclastic" of Wright and Bowes also appears unnecessary because it has no precise meaning ("other breccias") and simply refers to volcanic rocks formed beneath the surface. It is suggested that sub-volcanic is sufficient for this purpose.

Spilitic Fragmental Rocks

Autoclastic Breccias

Flow breccias, caused by disruption and deformation of flows and of pillows during extrusion, are common in the King Island lavas (particularly south of City of Melbourne Bay) and have already been described. Smaller scale breccias probably caused by disruption of congealed skins have also been described from Lynch Creek.

An intrusion breccia (sub-volcanic autoclastic) exists on the southern side of City of Melbourne Bay. Intrusion of spilite into dolomites and reddish shales has produced an irregular, discordant, plug-like mass of breccia consisting mainly of slabby fragments of creamy dolomite several tens of cm. long in a matrix of scoriaceous spilite. Several fragments of spilite up to 3 cm. diameter were also observed. The sediments and the breccia are overlain by a 15 m. bed of massive, coarse dolomite-spilite agglomerate that presumably was erupted through the vent.

Pyroclastic Breccias and Tuffs

The vitrophyres of King Island have already been described. Near the base of these volcanics just south of Conglomerate Creek there are several lenticular beds, only a metre or so thick, that consist of angular and subangular fragments of spilite and dolomite up to 20 or 30 cm. across. These breccias are probably formed by shallow explosions involving the dolomitic sediments immediately below the lavas. A thicker bed with coarser fragments overlies the intrusion breccia just described.

The presence on King Island of beds of agglomerate, many with dolomite fragments, at the base of the lava flows indicates that the main period of extrusion was preceded by a short phase of explosive activity. Minor flows and tuffs occur within the dolomitic sediments below the main spilite sequence.

Fine-to medium-grained basaltic tuffs occur at the base of the Crimson Creek formation just west of Zeehan (31847), west of Renison Bell (31845), and at other places. These rocks consist of varying amounts of albite, augite and spilitic rock fragments with subordinate quartz, iron oxides, chlorite etc. They are poorly sorted and the fragments vary from angular to somewhat rounded. They grade to tuffaceous sandstones derived from weathering of volcanics.

Andesitic Fragmental Rocks

Andesitic tuff outcrops on the Lake Margaret tram line west of Crown Hill (32526). It is crudely banded and granular in texture with more or less equidimensional grains (1/3-½ mm. diameter) of quartz, albite and hornblende and a little augite in a feldspathic fine-grained matrix. Albite predominates and again is remarkably fresh, showing typical, faintly brownish colouration in thin section. The hornblende is similar to that in the andesite lavas though the crystals have sharper and more regular margins; several crystals are fractured. Some calcite and epidote is also present.

Andesitic tuff of finer grain outcrops on the Tyndall track north of Crown Hill (610). It appears as a sand-grade (about 0.1 mm.)

aggregate of diopside grains and hornblende laths with a few quartz grains in an altered feldspathic matrix. Secondary epidote occurs as crudely developed granules.

Keratophyric Fragmental Rocks

Autoclastic and Pyroclastic Breccias

Many of the quartz keratophyres have a brecciated appearance on weathered surfaces but appear homogeneous on fresh faces and in thin section. The fragments vary from sub-angular to rounded and from microscopic to a metre or so in diameter. They are common in the Primrose Volcanics, in the Mt. Read Volcanics west of Tullah, at Lake Jukes and many other places. Thin sections show typical quartz keratophyre mineralogy and textures in the fragments. Autobrecciation is common in cooling rhyolitic magma (mainly because of its high viscosity) and flows and necks composed of rhyolite boulders and fragments in a rhyolite matrix are common (e.g. Williams, 1957). The unimodal Tasmanian breccias are typical, the majority being flows but some possibly representing vent fillings. The plug-like mass of breccia at Lake Jukes, containing angular boulders several feet across of Darwin keratophyre in a granophyric matrix, may be a vent filling.

A typical pyroclastic (?) breccia may be seen in Waterfall Gully (31272, 31273), on Whip Spur (32849, 31674), in the East Queen River (32835, 32852) (all near Queenstown), and on the Rosebery-Burnie road (30020). This rock is clearly fragmental and consists largely of angular to rounded fragments of pinkish and grey keratophyre up to 10 or 20 cm. diameter in a streaky feldspathic and chloritic base (Plate 30, No.1,2). The base consists of albite crystals, small rock fragments, ill-defined chloritic material and magnetite. None of the constituents of the groundmass appear waterworn and the degree of sorting is low. This rock type is probably an agglomerate. On Whip Spur the agglomerate overlies Darwin keratophyre and many of the rock fragments are pinkish in colur and similar microscopically to that keratophyre.

A rock (31857) of uncertain origin crops out on the Murchison Highway two miles north from Tullah (Plate 31, No. 1). It is made up almost entirely of keratophyric rock fragments, varying from 0.1 mm. to 30 or 40 cm. diameter, and arranged haphazardly in a chloritic and sericitic matrix in which weakly defined shared outlines are visible. Several of the rock fragments consist of a very fine-grained quartz-feldspar (?) aggregate partly replaced by untwinned (potassic ?) feldspar. The fine-grained base may well be devitrified glass which is commonly highly

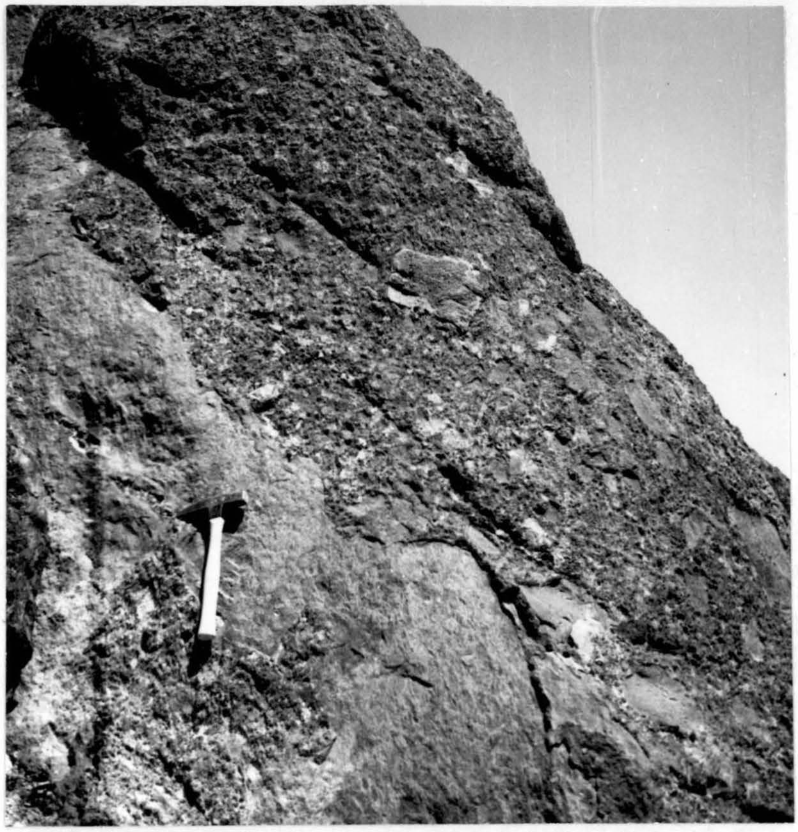


Plate 30 No. 1:- Agglomerate, Whip Spur.

Plate 30 No. 2:- Agglomerate, Waterfall Gully.
The width of the field is approximately 30 cm.

Plate 31 No. 1:- Agglomerate (?), Murchison Highway, two miles north of Tullah.

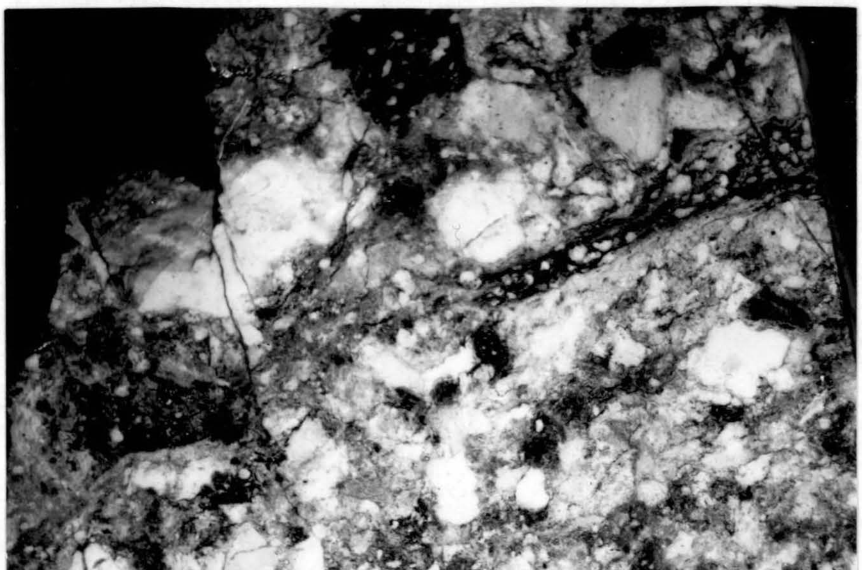


Plate 31 No. 2:- Agglomerate (?), Murchison Highway, north of Tullah. Natural size.

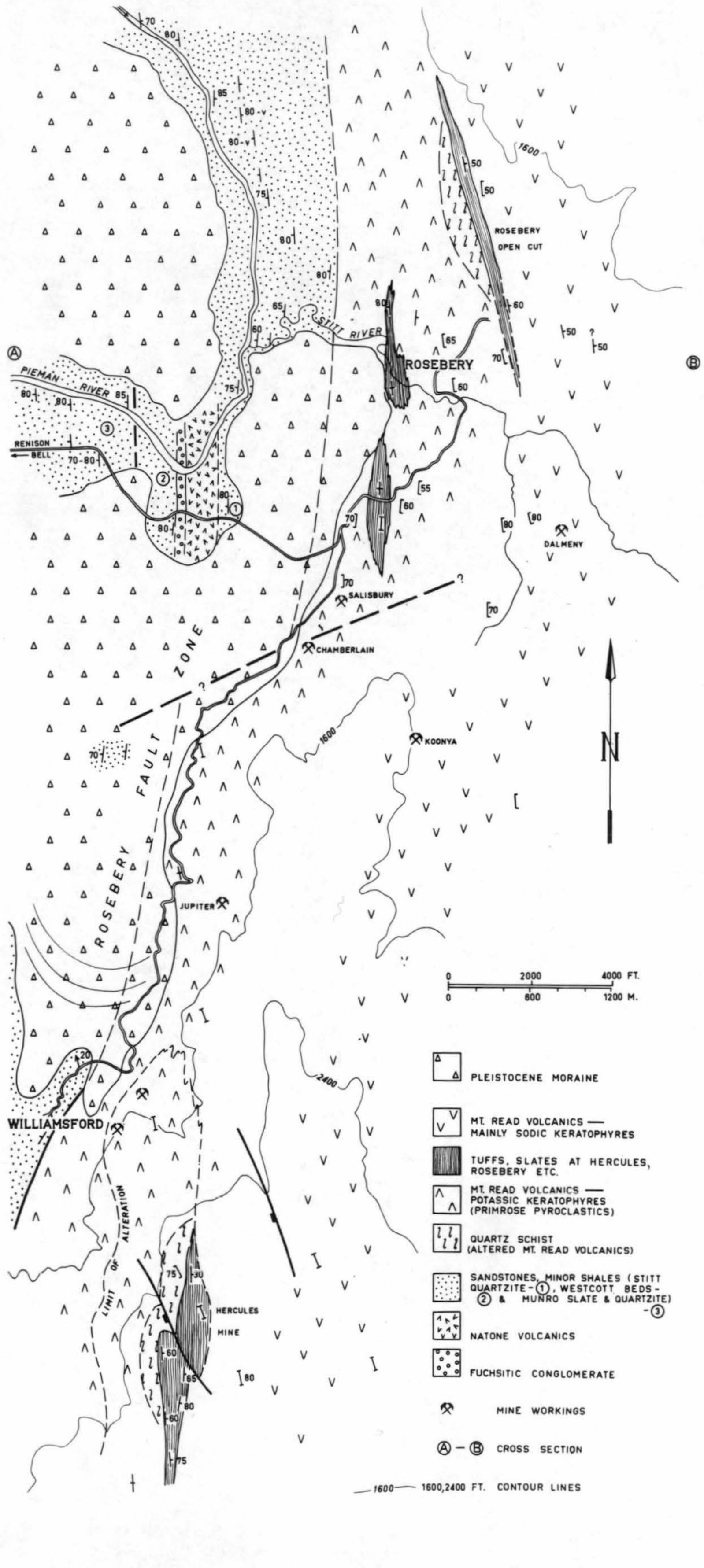
susceptible to replacement by K-feldspar. This rock may be an autoclastic breccia.

A little to the north of this outcrop is a rhyolitic breccia (31858) consisting mainly of angular fragments of pinkish or white quartz albite keratophyre together with fragments of dark green chloritic lava, all set haphazardly in a streaky chloritic matrix (Plate 31, No. 2). Many of the fragments have no albite or quartz phenocrysts and consist entirely of very fine-grained albite-quartz aggregates; presumably these represent quickly chilled, perhaps once glassy, fragments of quartz keratophyre. The texture of the rock is somewhat reminiscent of non-welded ignimbrites.

Potassic Breccias and Tuffs of the Primrose Volcanics and Similar Occurrences

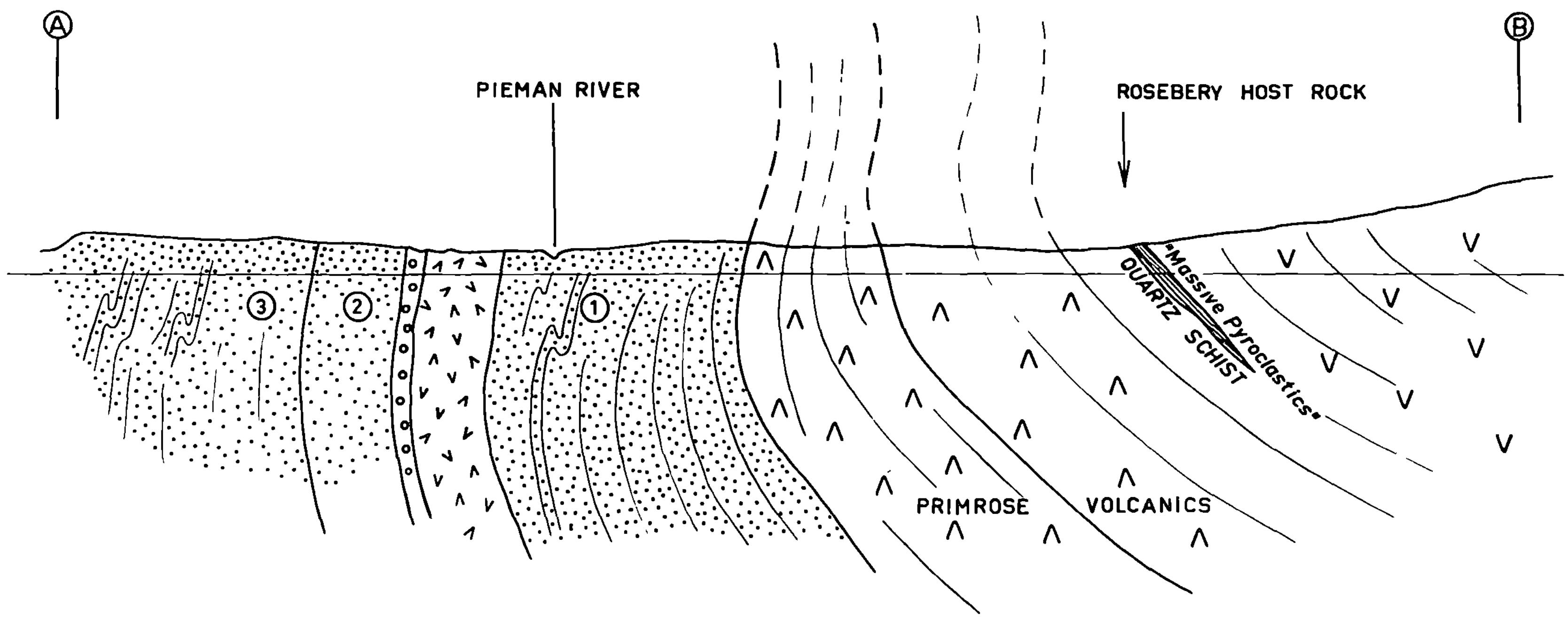
These rocks exposed between the Rosebery and Hercules mines (Figs. 11 and 12) were described as volcanics by Twelvetrees and Petterd (1899) and Hills (1923) but were regarded as intrusive by Finucane (1932) and Dallwitz (1946). They are characterised by their inhomogeneity which may be expressed as a faint, wispy banding (e.g. near the Williamsford turn-off), a blotchy appearance,

Fig. 11: Geological Map of the Rosebery-Hercules Area.



ROSEBERY
WILLIAMSFORD
STITT RIVER
PIEMAN RIVER
RENISON BELL
ROSEBERY OPEN CUT
DALMENY
SALISBURY
CHAMBERLAIN
KOONYA
JUPITER
HERCULES MINE
ROSEBERY FAULT ZONE
LIMIT OF ALTERATION
N
0 2000 4000 FT.
0 600 1200 M.
PLEISTOCENE MORAINE
MT. READ VOLCANICS — MAINLY SODIC KERATOPHYRES
TUFFS, SLATES AT HERCULES, ROSEBERY ETC.
MT. READ VOLCANICS — POTASSIC KERATOPHYRES (PRIMROSE PYROCLASTICS)
QUARTZ SCHIST (ALTERED MT. READ VOLCANICS)
SANDSTONES, MINOR SHALES (STITT QUARTZITE - 1, WESTCOTT BEDS - 2 & MUNRO SLATE & QUARTZITE) - 3
NATONE VOLCANICS
FUCHSITIC CONGLOMERATE
MINE WORKINGS
A — B CROSS SECTION
1600 — 1600, 2400 FT. CONTOUR LINES

Fig. 12: East-west cross-section through Rosebery, drawn from Fig. 11.


A
B
PIEMAN RIVER
ROSEBERY HOST ROCK
3
2
1
"Massive Pyroclastics"
QUARTZ SCHIST
PRIMROSE
VOLCANICS

or pronounced breccia texture. The breccias (exposed, for example, near the Rosebery Staff Tennis Courts) contain a high proportion of spherical, slabby or angular fragments of white quartz porphyry from several cm. to a few mm. across and there are up to 20% of streaked-out lenses of greenish-grey pumice (?) or glassy material. All these constituents tend towards parallel orientation to give rocks closely resembling coarse welded tuffs or breccias. The brecciation and streaming are generally unrelated to cleavage which is only weakly developed in most of these rocks. The volcanics contain lenticular beds of finely banded tuffs, the uniform banding suggesting a depositional origin.

Microscopically, most of the Volcanics consist of feldspar and corroded quartz crystals in a base that is largely quartz and sericite. Both albite and K-feldspar occur as "phenocrysts". The base in many rocks is featureless but vitroclastic texture is revealed in some (e.g. 32924, 32952 and in Plate 32, Nos. 1 and 2, and Plate 33). In Plate 32, No. 1 the albite crystals are fractured and broken. In the other two plates, both of core from drill hole H391 (Hercules), considerable deformation of the shards is evident (distinct from cleavage deformation) and the rocks closely resemble welded tuffs with eutaxitic texture. Fragments of collapsed pumice may be seen in Plate 33.

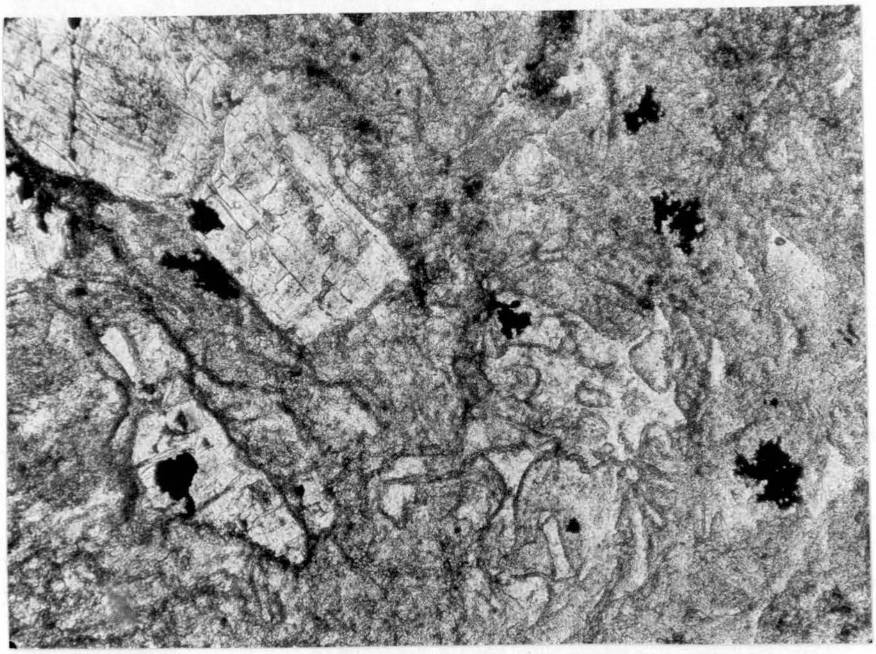


Plate 32 No. 1:- Fractured albite crystal in shard-rich matrix, north of Hercules Mine, X65.

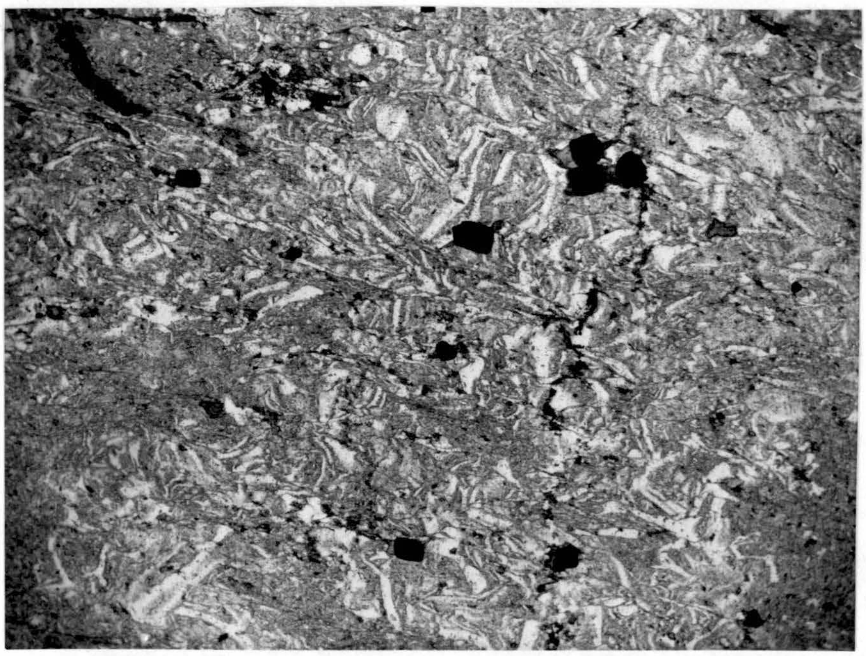


Plate 32 No. 2:- Deformed shards in Tuff, from drill hole H391 (Hercules) at 587 ft. (32954). Note crudely developed cleavage. X25.

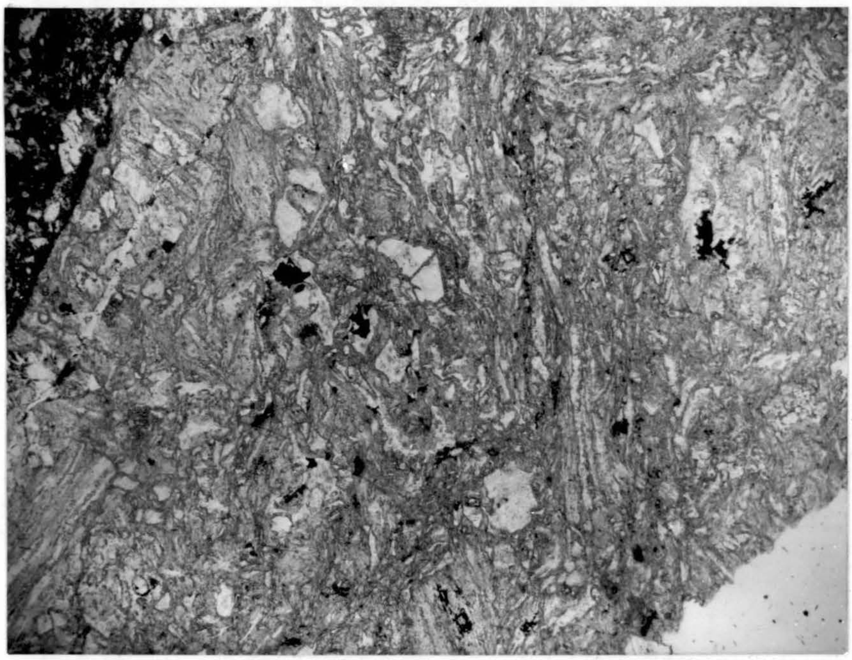


Plate 33 :- Deformed shards, phenocrysts and pumic fragments in welded tuff (?), from drill hole H391 (Hercules) at 270 ft. (32955). X25.

A characteristic of these rocks is the presence of euhedral albite laths with rims of K-feldspar. The rims vary in thickness (both between rims and within rims), reaching a maximum measured width of 0.1 mm. The inner margin is sharp against albite but the outer margin is highly irregular or scalloped (Plate 34, Nos. 1 and 2), the surrounding material being generally a fine-grained albite or an albite and K-feldspar aggregate. The K-feldspar in some rims is homogeneous and in one showed Carlsbad twinning; in others it shows fine lamellar structure (twinning ?) of rather patchy and uneven distribution and some crystals show fine "grid-iron" twinning. The rim is in crystallographic continuity with the albite core (as shown for instance by the common basal cleavage - see also Battey, 1955) and as a result the optical axes are at a small angle, causing the rim to extinguish a few degrees off the albite extinction position. In one slide (31789A) a rimmed albite crystal encloses a K-feldspar crystal which has the same crystallographic orientation as the albite and the same optical orientation as a thin rim of K-feldspar around the albite margin.

The rims show slight penetration along the cleavage planes of the albite cores and irregular patches of K-feldspar occur within the albite. In some cases veinlets extend across the core;

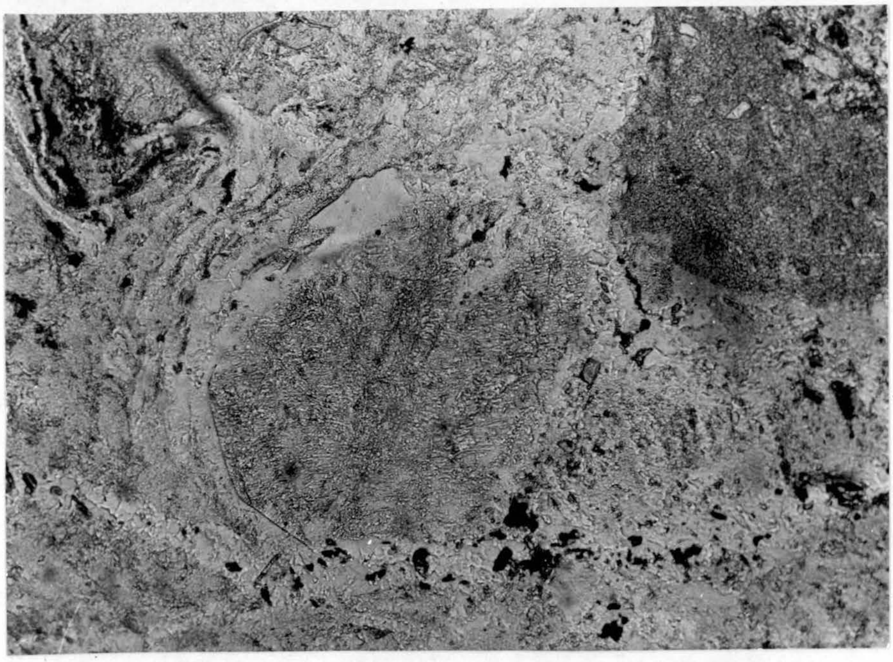


Plate 34 No. 1:- Albite crystal with rim of clear K-feldspar, near Rosebery (32953) ordinary light. X65.

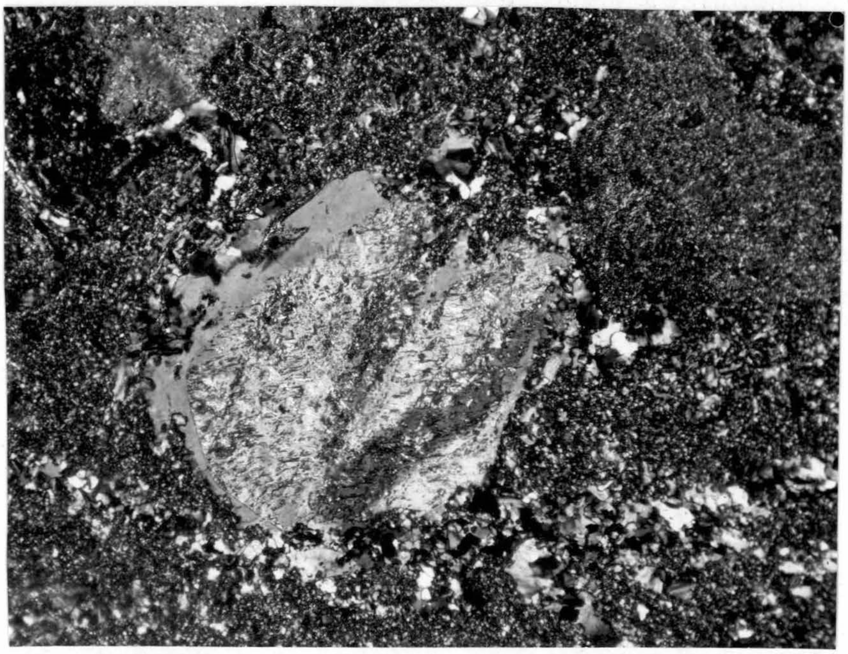


Plate 34 No. 2:- Albite crystal with rim of clear K-feldspar, near Rosebery, crossed nicols.

these are crudely sub-parallel but do not appear to have any special relation to the crystallographic direction of the host (Plate 36, No. 1). In some they are stepped by local orientation along cleavage or they may follow cleavage lines for most of their length. In 32923 an albite crystal has been slightly displaced along the K-feldspar veinlets, indicating some movement before K-feldspar development.

Most of the Primrose Volcanics contain individual crystals of K-feldspar (Plate 35, Nos. 1 and 2), in some rocks to the exclusion of albite (32924, 32925). The K-feldspars are up to 1 x ½ mm. in size, are anhedral, very clear and generally show patchy extinction (Plate 34, No. 2 and 35 No. 2) and a fine, rather uneven grid-iron twinning (?) like that in the rims; the composition planes are (010) and approximately (100), apparently similar to microcline twinning. In some cases thin lenses of albite (?) are aligned parallel to (100), indicating the presence of perthite. In others, almost rectangular areas of albite are arranged through the potash host, the margins tending to alignment along (010) and the (100) planes, giving a crude checkerboard texture.

Finucane (1932) quoted several analyses of the Primrose Volcanics and these are reproduced in Table 11. As already mentioned, these rocks are partly altered and possibly metasomatised.

Table 11: Primrose Volcanics and other Pyroclastics

	1	2	3	4	5	6
SiO_2	81.12	73.76	70.00	70.00	69.20	73.89
TiO_2	0.07	0.05	0.22	0.25	0.27	0.22
Al_2O_3	13.27	14.80	17.81	17.35	15.67	15.02
Fe_2O_3	0.29	4.08	1.86	0.93	0.43	1.28
FeO	0.77	0.97	2.45	2.77	3.09	1.15
MgO	0.36	0.58	1.45	0.65	0.79	1.45
CaO	Tr.	0.10	0.15	1.48	1.50	0.40
Na_2O	0.83	0.33	1.65	2.58	2.22	4.06
K_2O	3.15	4.83	2.92	4.15	4.47	1.43
H_2O+						
H_2O-						
MnO					0.37	
P_2O_5	0.042	0.01	0.025	0.075	0.053	
CO_2						
FeS_2						
S	0.006	Tr.	Tr.	0.068	Tr.	
Ig. loss	1.80	2.30	2.50	1.50	3.60	1.44
Total	101.708	101.81	101.035	101.803	101.663	100.25

Table 11: Primrose Volcanics and other Pyroclastics (continued)

		1	2	3	4	5	6
qu		63.00	52.38	45.66	33.42	32.21	41.75
or		18.90	28.32	17.21	24.52	26.67	8.34
ab		6.82	2.62	14.15	21.83	18.85	34.05
an			0.56	0.66	6.82	7.23	1.95
cor		8.47	9.07	11.70	6.11	4.49	6.12
hy	en	0.90	1.45	3.60	1.62	2.00	3.60
	fer	1.06		2.51	3.90	5.55	0.66
mag		0.46	2.78	2.78	1.35	0.70	1.90
hem			2.24				
calc							
ap				0.06	0.19		
ilmen		0.15	0.15	0.46	0.47	0.46	0.45
Total		99.77	99.57	98.79	100.23	98.28	98.82

Table 11: Primrose Volcanics and other Pyroclastics (continued)

	7	8	9	10	11
SiO_2	75.73	71.25	68.88	58.6	69.48
TiO_2		0.37	0.47	n. dt.	0.65
Al_2O_3	12.70	14.06	13.47	15.7	14.04
Fe_2O_3	2.25	1.16	0.59	8.9	1.33
FeO		2.73	2.32		3.50
MgO	0.60	0.73	1.29	3.6	2.15
CaO	2.00	1.19	2.75	2.8	0.60
Na_2O	3.48	3.29	5.40	5.8	0.24
K_2O	2.04	3.38	1.85	1.8	4.23
H_2O+		1.47	0.84	1.0	2.79
H_2O-		0.26	0.18		Nil
MnO		0.03	0.12	n. dt.	Tr.
P_2O_5		0.15	0.14	n. dt.	0.09
CO_2		0.01	1.89	0.7	
FeS_2					0.91
S		0.38	0.14		
Ig. loss	1.20				
Total	100.00	100.46	100.56	98.9	99.99

Table 11: Primrose Volcanics and other Pyroclastics (continued)

		7	8	9	10
qu		40.62	34.44	26.66	1.08
or		12.06	20.02	10.93	10.64
ab		29.45	27.77	45.69	49.08
an		9.92	5.00	0.78	9.47
cor		1.13	3.16	2.30	0.74
hy	en	1.49	1.80	3.21	8.97
	fer	4.14	3.43	3.22	16.34
mag			1.62	0.86	0.00
hem					
calc				4.30	1.59
ap			0.34	0.33	0.00
ilmen			0.76	0.89	0.00
Total		98.81	98.34	99.17	97.91

Table 11: Primrose Volcanics and other Pyroclastics

1-6. Quartz keratophyres of the Primrose Volcanics, Rosebery-Hercules area (from Finucane, 1932).

1. Williamsford Road, near Williamsford - R176.
2. Quarry near staff tennis court (Barkers Crossing) - R1.
3. Quarry a few chains east of Primrose No. 2 tunnel - R11.
4. Quarry at corner of Primrose and Dalmeny Streets - R177.
5. Ridge near NW corner of Dalmeny Lease (south of Rosebery) - R97.
6. Sodic keratophyre, west side of slates 2½ chains (50 m.) north of No. 1 adit, Rosebery Mine - R265.

7. Sodic keratophyre, North Hercules Section, Mt. Read (Twelvetrees and Petterd, 1898, p. 46).

8. Primrose Pyroclastics, 500 m. north of Hercules Mine, specimen 31789A. Analyst: Japan Analytical Chemistry Research Institute.

9. Quartz-feldspar crystal tuff, Lynch Creek, specimen 31657. Analyst: Japan Analytical Chemistry Research Institute.

10. Albite-augite tuff, Lynch Creek specimen (the augite trachyte (?) of Solomon, 1960).

11. Banded siltstones and tuffs (31631), South Owen Creek. Analyst: Dept. of Mines, Tasmania.

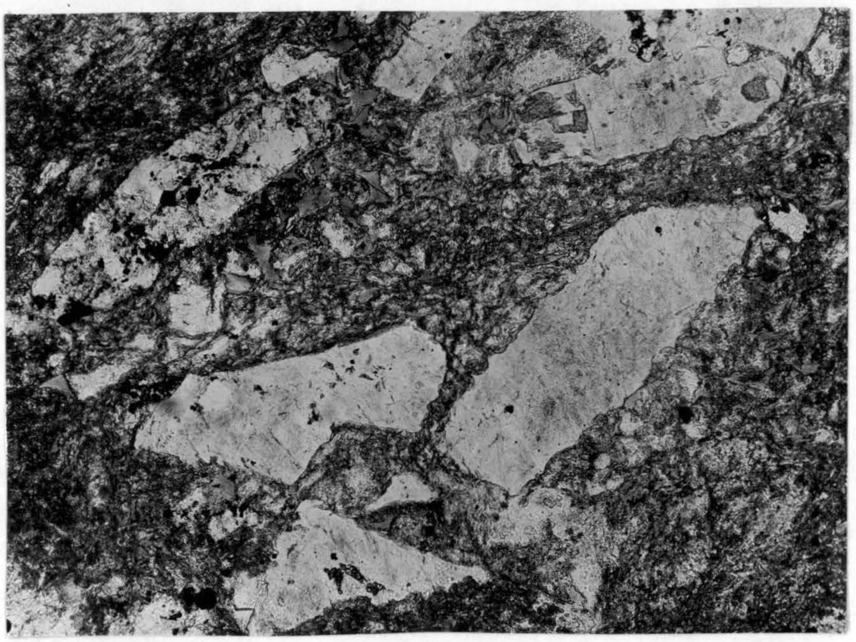


Plate 35 No. 1:- K-feldspar phenocrysts in shard matrix, near Rosebery, ordinary light. X45.

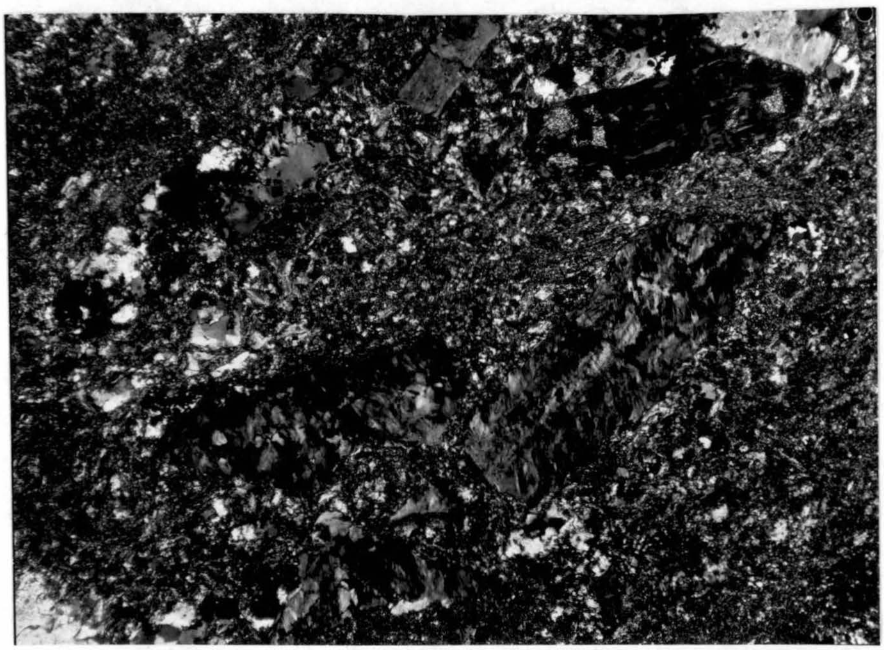


Plate 35 No. 2:- K-feldspar phenocrysts in shard matrix, near Rosebery, crossed nicols.

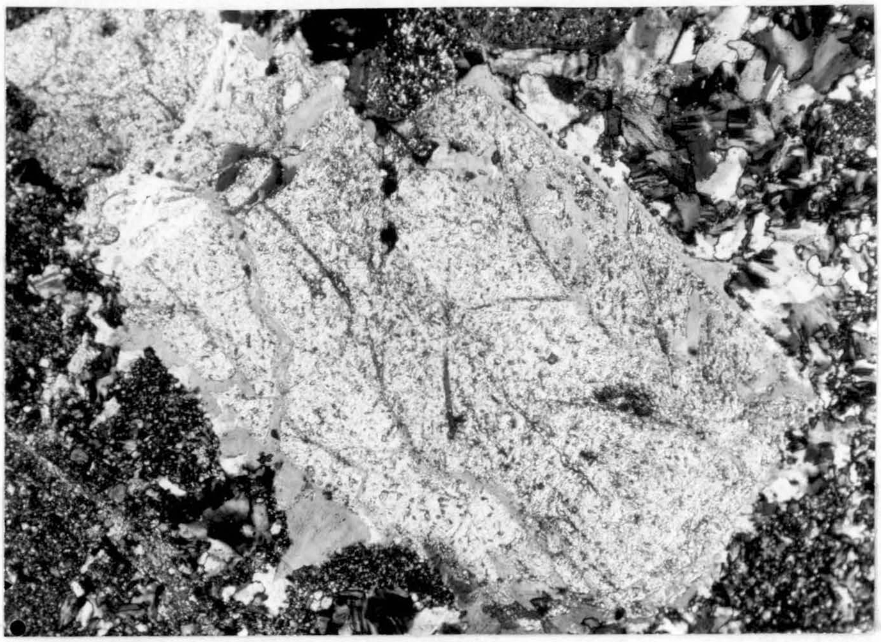


Plate 36 No. 1:- Altered albite with K-feldspar rims, patches and veinlets, near Rosebery (32953). X70.

Plate 36 No. 2:- Pumice froth on albite crystals, Hercules mine (50G161) ordinary light. X5.

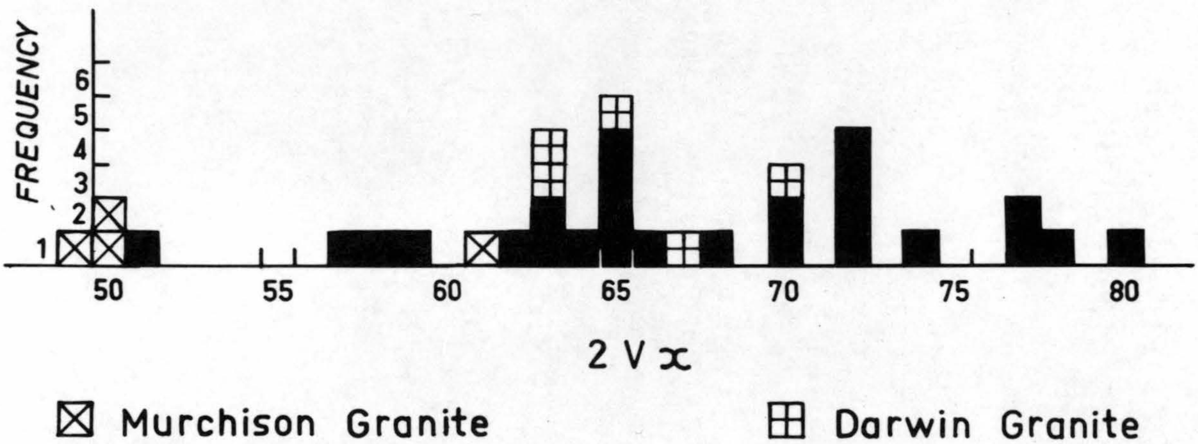
Similar rocks have been found in holes drilled by the Department of Mines in search of barite at Madam Howard Plains, 3 miles north-west of Queenstown (Mines Department 63-278 to 63-283). They are porphyritic and characterised by albite crystals with rims and veinlets of K-feldspar in a fragmental but recrystallised matrix. They cannot be definitely classed as a quartz keratophyre composition as no analyses are available but quartz occurs as small crystals and also is fairly plentiful in the feldspar-quartz aggregate that makes up the groundmass. No ferromagnesian minerals are present. Apatite laths are present, up to 1.5 x 0.1 mm. in size (Mines Department 63-279), and calcite and sericite probably represent the products of late alteration. The feldspar-quartz matrix varies considerably in grain size and is characterised by irregular coarser patches that appear to have been recrystallised. The albite crystals are up to 1 x 1 mm. in size and have K-feldspar rims which may be as wide as 0.1 mm. in places, as noted in the Primrose Volcanics; the inner rim margin is sharp against the albite but irregular against the groundmass. K-feldspar also occurs in veinlets apparently replacing albite crystals. Blebs of albite may be seen in some K-feldspar rims.

The <u>composition of the K-feldspar</u> in the Primrose and

Madam Howard rocks is uncertain as no refractive index measurements were possible. It differs optically from that in the Darwin keratophyre in being particularly clear and generally possessing grid-iron twinning but may well have a similar composition; the spread of optic axial angles, 57° to 78°, is only slightly less than that for the feldspars in the Darwin keratophyre. Fig. 13 shows the distribution of 2V in the K-feldspars of the Primrose Volcanics and Darwin keratophyre.

Several specimens from the Hercules Mine (618, 619, 50G-161 and 162, 32807) are composed almost entirely of randomly oriented pumice fragments. Fragments up to 3 mm. long are completely collapsed and streaked out and frayed, but larger fragments consist of contorted, streaky pumice with only slightly flattened or even undisturbed gas holes. Under crossed nicols, these features are almost invisible and the rocks appear as heterogeneous, very fine-grained quartz-feldspar aggregate with a little chlorite outlining the textural boundaries. In 50G-161 and-162, and from borehole H391 (at 293 ft.-32956), the rock consists of lumps of frothy pumice centred on altered albite crystals (Plate 36, No. 2). The albite crystals have thin K-feldspar rims and the pumiceous surround, several mm. wide, is composed of an albite and K-feldspar aggregate. It shows distinct circular structures

Fig. 13: Optical axial angle measurements for all K-feldspars in the Mt. Read Volcanics and the Darwin and Murchison Granites.


FREQUENCY
1
2
3
4
5
6
50
55
60
65
70
75
80
2 V x
Murchison Granite
Darwin Granite

(Plate 37, No. 1) composed of albite and outlined by chlorite films. Similar vesicular (?) rims to albites occur in the Madam Howard Plains rocks (e.g. Mines Department, 63-284). These rims probably formed by adhesion of gas bubbles to crystals floating in molten lava and may have developed after extrusion or in the vent. The bubbles probably formed a sheath or "froth" around the crystal nuclei, causing the crystals to rise and form crusts to the flows.

In Plate 37, No. 2 discrete zones of flowage can be seen crossing the froth and these must have developed by movement in the froth crust and injection of still molten material from below.

Collapsed and non-collapsed pumice fragments are common in 596, 597, 600, 676 and in specimens from borehole R142 (e.g. at 160 ft.), from the Rosebery Mine. These occur with altered albite crystals in a matrix of slightly deformed or non-deformed shards.

Concerning the presence of collapsed pumice fragments in non-welded tuff matrix, Ross and Smith (1960) suggested that this was due to differential compaction related to the relatively lower viscosity of the volatile-rich pumice compared to the glassy matrix. However, Fiske (1963) figured stretched, "long-tube" pumice fragments in his pyroclastic flows and suggested (p. 399) that "the pumice was

Plate 37 No. 1:- Froth rim to albite crystals, Hercules mine (32956), ordinary light. X45.

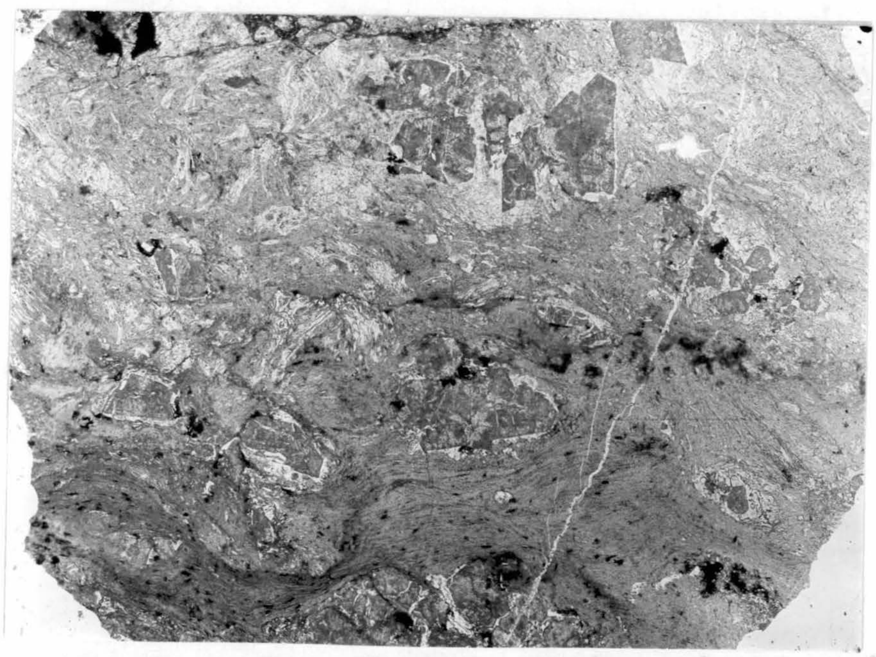


Plate 37 No. 2:- Flowage zones in pumiceous rock, Hercules mine (50G162), ordinary light. X5.

stretched as it was extruded and was later mixed with undeformed glass shards and glass dust.....".

Origin of the Primrose Volcanics

Determination of the process forming fragmental acid volcanics of this type is difficult, particularly in such old and metasomatised examples as the Primrose Volcanics.

Several specimens have the appearance of welded tuffs (see, for example, Ross and Smith, 1960 and Martin, 1959) and the rocks compare closely in hand specimen and thin section to samples from the Waiotapu, Ongatiti, Ahuroa and Whakamaru ignimbrites (kindly sent to the writer by Professor Bradley and numbered 30877 to 30886).

Non-welded tuffs are difficult to identify positively and welding on all scales (with formation of eutaxitic structure) appears to be the most significant indication in areas of old rocks where field relationships may not be too clear (e.g. Oliver, 1954 on Wales and the English Lake District, and Beavon et al., 1961 on Snowdonia).

The possible presence of ignimbrites on the West Coast was first hinted at by the writer (1960) at an early stage of this study, and Spry (1962b, p. 259-260) suggested the Primrose Volcanics might be ignimbrites. Campana and King (1963, p. 19) believed that "the presence of true ignimbrites appears unquestionable though no evidence had been given to that time to support their own or any earlier suggestions.

The Primros e Volcanics are particularly variable though the majority are fragmental and/or pumiceous, and holocrystalline types are rare. They may be largely welded and non-welded ignimbrites derived from nuées ardentes or be in part more closely related to "froth flows".

McCall (1962), Boyd (1961), and Kennedy (1955, p. 495) believe welding in acid fragmental rocks can be due to flowage of highly gaseous lava and not necessarily to compaction in a nuée ardente deposit. McCall, describing single but composite flows from Kenya which contain crystalline, eutaxitic and pumice-froth phases, believed these different products may be explained by combinations of "brittle" and "plastic" behaviour after eruption. Parts of the erupted material crystallise in the normal way to produce essentially crystalline phases. Other parts vesiculate freely to form an expanded lava froth ("froth flow") in which streaming and shredding of the plastic material during flows may produce a welded-like texture. Disruption by explosion gives rise to fragmental phases containing pumice shards and crystal fragments.

The presence of fragments of crystalline keratophyre in the streaked out devitrified breccias at Rosebery seems a natural product of a froth flow with crystalline phases, rather than a nuee ardente.

Ignimbrites and froth flows described to date are sub-aerial and an air medium seem to be regarded as essential for ignimbrites (see Rankin, 1960). However, there seems to be no reason why an ignimbrite should not develop under water and travel as a discrete unit enveloped in steam and other gases provided it is charged with sufficient solid matter to make it substantially heavier than water. An aqueous environment is indicated for the Primrose Volcanics by the finely banded tuffs and siltstones that occur above and to some extent within the Primrose Volcanics (and the Madam Howard rocks) but they may have formed in local lakes or lagoons alongside and within volcanic ranges. It is certainly difficult to imagine pumice froths and disrupted pumiceous material remaining under water because of the low density of the pumice. In a succeeding section are described some crystal-vitric tuffs from Teepookana that were deposited in an aqueous environment but these lie some distance west of the Mt. Read Arc and in the centre of the Cambrian marine basin.

Natone Volcanics

These volcanics crop out west of Rosebery and appear to be older than the Primrose Volcanics (Fig. 11). Specimens (31793A, 31794A, 32863, 32903) vary from cream to medium green-grey and are mostly inhomogeneous, showing streaks or lenses of fine-grained, darker material. Crystals of feldspar up to 3 mm. diameter, and forming 10-20% of the rock, are visible in some specimens. In thin section the rocks are largely composed of a micro-crystalline aggregate showing polygonal cracking, with a few albite crystals and strongly cracked quartz crystals. Mild sericitisation is general and differential sericitisation highlights textures in the matrix in which wisps and shards of once-glassy material are prominent (Plate 38, No. 1). The shards may be somewhat drawn out along a Devonian cleavage.

The Natone Volcanics do not show features characteristic of ignimbrites and appear to be mainly vitric tuffs. They lie between cross-bedded sandstone (the Stitt Quartzite) and finely banded mudstones and sandstones and may be of aqueous origin.

Crystal-Vitric Tuffs of Teepookana

Somewhat similar volcanics occur several miles west of the Mt. Read Volcanic Arc, in the King River gorge, two miles

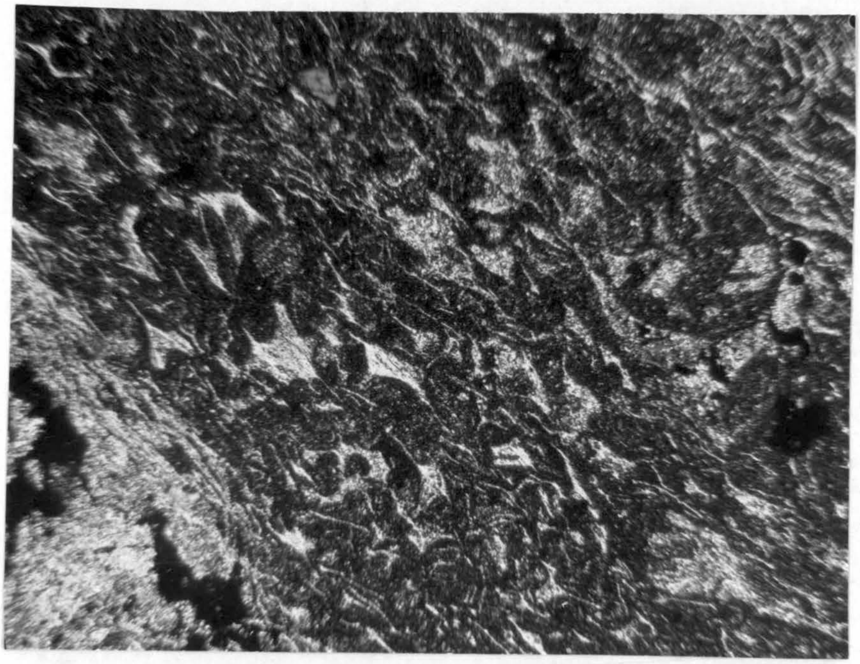


Plate 38 No. 1:- Typical texture seen under crossed nicols in Natone Volcanics. X45.

Plate 38 No. 2:- Frayed end to pumice fragment, Teepookana. X115.

upstream from the mouth and near Teepookana (Fig. 3). Several hundred feet of volcanics are exposed on the north bank of the river and the section is approximately as follows (youngest first):

(1) 2 m.: Medium grey, coarse feldspathic tuff, regularly jointed (31772A).

(2) 1.2m.: Dark-medium grey siliceous shale with poorly defined cleavage.

(3) 2 m.: Similar to coarse tuff above but more coarsely grained.

(4) 4.5m.: Dark grey, crudely banded siliceous shale with spheres of pyrite scattered throughout (31773A, 31770A, 31757A).

(5) 6 m.: Pale-medium grey, aphanitic fine-grained tuff, regularly jointed. More coarsely grained (sand grade) tuff in places; lower 2 m. shaley and dark grey.

(6) 3 m.: Very pale grey cherty tuff with subconchoidal fracture (31771A, 32850).

(7) 6 m.: Similar to fine-grained tuff of (5).

(8) 12 m.: Dark grey siliceous shales and pale or dark-medium grey chert (31762A) in fine bands (31759A); local intraformational crumpling.

(9) 60 m.: Pale grey, fine-grained tuff, as for (5) with crude banding on some weathered surfaces, and coarse-grained feldspathic tuff (31774A, 31775A, 31776A, 31756-58).

(10) 60 m.: Grey to fawn-coloured, coarse-to fine-grained feldspathic tuffs with occasional banded chert layers a few cm. thick.

(11) 6 m.: Very pale grey, almost creamy, feldspathic tuff or lava (31842).

(12) approx. 30 m.: Feldspathic coarse tuffs (31778A), siliceous pyritic shales, and fine tuffs. The tuff may contain dark green slivers up to 1 cm. long.

East of this section and possibly faulted against it, are 50 metres or so of similar lavas, tuffs and sediments. Conspicuous among these are two beds, 3 to 5 m. thick, of uniform, medium-grey, fine-grained tuff (31753A, 31754A, 31769) separated, and also overlain, by a metre or so of dark shale. These tuffs are similar to beds (5) and (7).

Apart from the black shales of (2) and (4) the succession consists entirely of vitric and crystal tuffs, though all the glassy material is now devitrified. Beds (1) and (3) consist of a poorly sorted, randomly arranged aggregate of albite laths and fractured quartz fragments up to ½ mm. long in a dark, dusty matrix.

Beds (5), (7), (9) and part of (12) are mainly homogeneous, fairly massive units with a rough sandy feel. They consist of a rather murky, microcrystalline, partly sericitic matrix with or without fragments of fractured quartz crystals up to 0.02 mm. diameter scattered randomly; these fragments appear to be parts of quartz phenocrysts. Conspicuous in places are wispy slivers and blebs of pyrite. The rock was probably originally largely glass. 31754A shows drawn-out blebs of devitrified (?) material and 31760A has a faint banded or streaky pattern. Analyses are given in Table 12, Nos. 1 and 3.

The coarser varieties are highly feldspathic and poorly sorted wtih a disrupted framework, with randomly arranged stumpy feldspar laths up to 2 or 3 mm. in length, in a matrix varying from dark grey to pale grey and in places appearing streaky (31756A, 31758A). The feldspars are partly altered albite, many of the crystals having thin rims (0.03-0.04 mm. wide) of potash-feldspar identical to that described in the section on potassic quartz-keratophyres. The rims are in crystallographic continuity with the albite core, as evidenced by the common cleavage and near-coincidence of optical axes. The "groundmass" to the albite crystals is in part a ferruginous dust but in places it clearly contains fragments of banded, microcrystalline aggregate that are devitrified, collapsed pumice fragments (Plate 38, No. 2). They are arranged

Table 12: Teepookana Tuffs

	1	2	3	4
SiO_2	74.84	77.78	71.26	67.56
TiO_2	0.25	0.26	0.38	0.79
Al_2O_3	12.03	13.02	14.15	11.70
Fe_2O_3	0.99	0.07	0.99	0.55
FeO	1.60	1.09	2.30	3.84
MgO	2.12	0.49	0.83	2.35
CaO	1.10	0.50	4.00	1.06
Na_2O	0.90	6.04	0.76	1.27
K_2O	3.70	0.30	3.47	2.63
H_2O+	2.73	0.71	2.10	3.30
H_2O-	0.27	0.09	0.10	0.40
MnO	0.09	0.02	0.03	1.13
P_2O_5	0.07	0.20	0.17	0.34
CO_2	Nil			
FeS_2				2.39
SO_3				0.65
Total	100.69	100.57	100.54	99.96

<u>Table 12: Teepookana Tuffs</u> (continued)

qu	49.27	39.56	42.94
or	21.87	1.77	20.51
ab	7.62	51.11	6.83
an	5.00	1.17	18.73
cor	4.71	2.33	2.28
hy en	5.28	1.22	2.07
hy fer	1.88	1.55	2.83
mag	1.44	0.10	1.44
ap	0.16	0.47	0.40
ilmen	0.48	0.49	0.72
Total	97.71	99.77	98.75

1. Fine-grained feldspathic tuff (?), specimen 31769, west of Teepookana, King River.
2. Pale cherty tuff (?), specimen 31771A, same locality.
3. Fine-grained feldspathic tuff (?), specimen 31775A, same locality.
4. Black pyritic shale, same locality. A "small amount" of organic matter is reported by the analyst. Specimen 31770A.

Analyst: 1-4: Department of Mines, Tasmania.

at random and a few are folded about crystals. The fragments have frayed ends (Plate 38, No. 2) typical of compressed pumice fragments in ignimbrites (see, for example, Martin, 1959).

Specimen 31758A likewise consists of a random arrangement of crystals, pumice, fine-grained lava fragments in a ferruginous, cryptocrystalline dust but it also contains flattened ovoids consisting of almost "fibrous" albite arranged perpendicular to the walls of the ovoids. The fibres from adjacent walls may meet at a central crack or a ferruginous zone (Plate 39, Nos. 1 and 2) or may continue right across the ovoid as single crystals. These bodies are very much like the partly devitrified pumice fragments illustrated by Ross and Smith (1960, Figs. 57, 58) and believed to be characteristic of ignimbrites. In the relatively young examples these authors describe the fibres are cristobalite-feldspar intergrowths and develop by devitrification of the glassy pumice. The pumice ovoids in the Teepookana tuffs have no preferred orientation.

Part of one tuff horizon shows a crude, very fine lamination which is seen under the microscope as a streakiness in a dusty, very fine-grained matrix.

Beds (6) and (11) consist of creamy "chert" traversed by quartz veinlets and possessing a characteristic almost conchoidal, fracture. The "chert" is a microcrystalline ($<$ 0.001 mm.

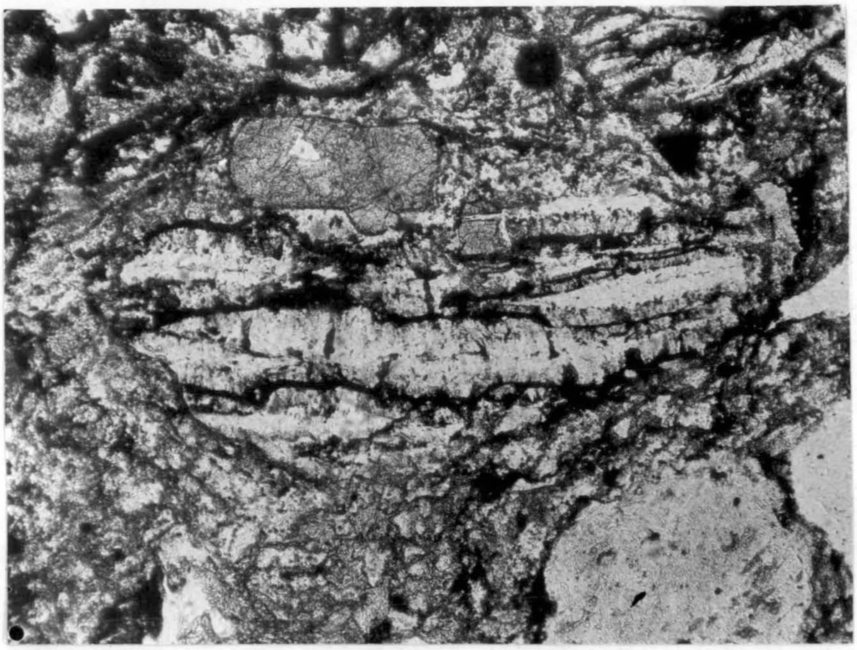


Plate 39 No. 1:- Pumice ovoid in tuff, Teepookana, ordinary light. X65. At top centre is an apatite crystal and at bottom right an albite crystal.

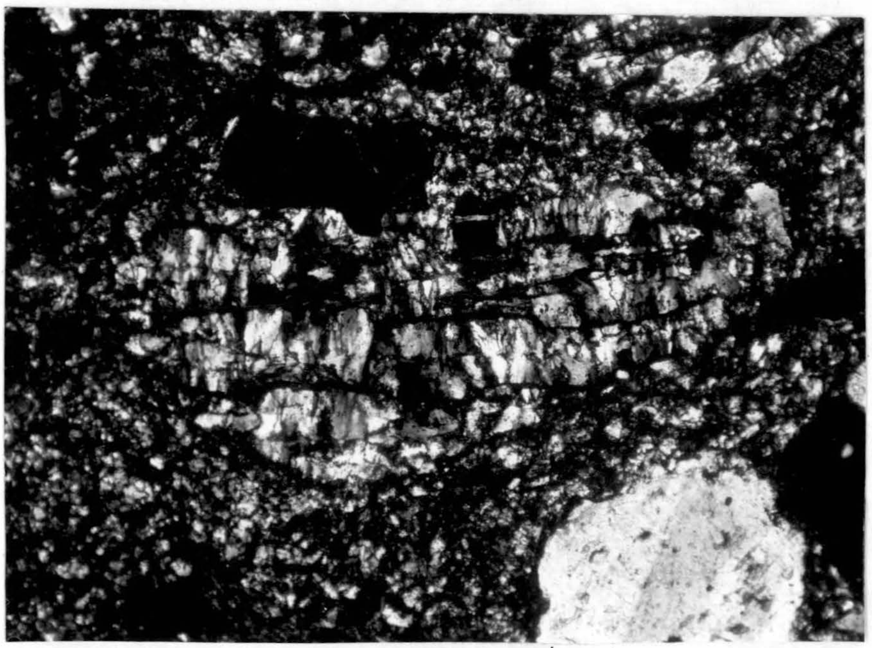


Plate 39 No. 2:- Pumice ovoid in tuff, Teepookana, crossed nicols.

diameter) aggregate not dissimilar to the apparently devitrified glass of beds (5) and (7), though it is finer in grain. The highest refractive index of the matrix aggregate is 1.538 ($\pm$ .001) and it gives X-ray patterns indicating that it is mainly quartz and albite; the chemical analysis of Table 12 (No. 2) confirms this identification. Parts of the material contain slivers, lenses and subparallel bands of darkish grey, streaky material. In one specimen the top part of a layer several cm. thick of alternating dark and pale grey material appears to have been crumpled and streamed out as slivers and blebs in the pale chert. Microscopically, it is similar to the pale tuff but has a crude banding with pale and dark grey material; it is probably a devitrified vitric tuff.

Some of the cherty material is a darkish grey colour and has pronounced conchoidal fracture - rib and hackle fractures (Solomon and Hill, 1962) are common on broken surfaces. In thin section, a pronounced polygonal cracking is visible (Plate 40) in a cryptocrystalline, dusty matrix that is probably devitrified glass (31762A). X-ray diffractograms indicate quartz, albite and orthoclase. In 31778A, a similar matrix is ramified and replaced by an irregular network of sericite and calcite, the remaining patches of unaltered material showing polygonal cracks. The alteration gives the rock a whitish colour in hand specimen and it has

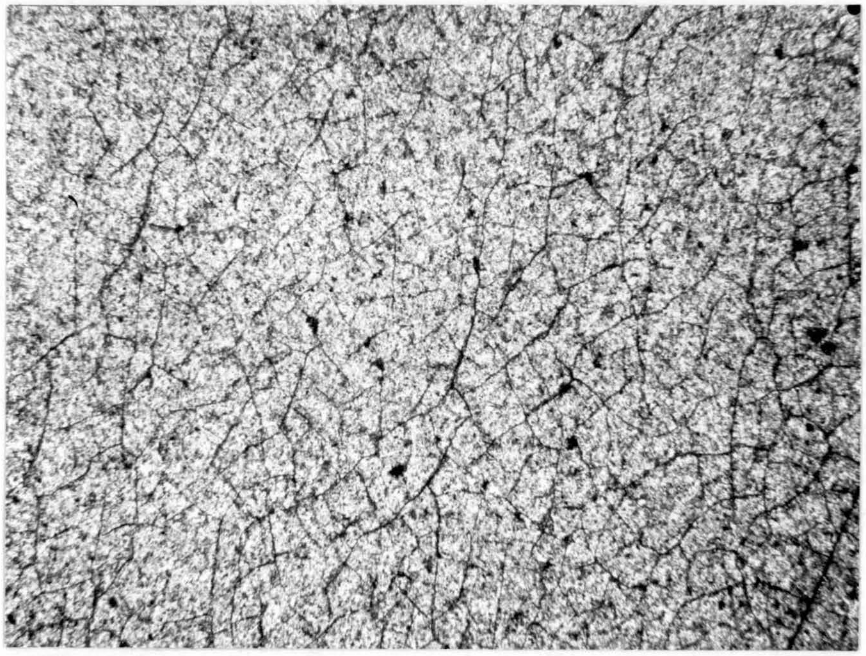


Plate 40 :- Polygonal cracking in matrix, Teepookana.

X 45.

lost its conchoidal fracture. Fragments of dark material up to ½ cm. wide are visible and these are probably once-glassy fragments. Similar altered material occurs along strike near the 21 mile peg on the Strahan-Queenstown road (31797A); this rock has a fine speckled appearance (due to calcite crystals), is faintly streaky and has a pale greenish tinge due to the presence of sericite.

Bed (11), a milky grey, cherty rock, is the only layer that looks like a lave in thin section. It consists of 10% or so of albite laths up to 1/3 mm. long in a fine-grained, interlocking aggregate of albite and possibly orthoclase. Some of the albite crystals have thin rims of K-feldspar. Distinction between fine-grained lavas and tuffs is extremely difficult and several errors in identifying these rocks may well have occurred.

Fine-grained, vitric (?) tuffs similar to those of the Teepookana area may be seen at several points on the Zeehan-Queenstown road between the Henty and Yolande Rivers (31716A, 31728A, 31719A, 31730A, 31731A, 31735A, 31764A). Some of these show partial replacement by calcite and sericite. Some of the specimens (e.g. 31730A) are vitric-crystal tuffs with fragments of albite crystals and a little quartz scattered through the matrix. 31731A contains approximately 40% albite crystals up to 0.5 mm.

long, a few percent of lava fragments and 10% of distorted pumice fragments of similar size. The rock differs from the coarse Teepookana tuffs in that the pumice fragments are numerous and are considerably distended and contorted. In 31741A (from near the Henty River bridge) pumice fragments with characteristic frayed ends are numerous and appear to have been considerably distended and locally intruded into the matrix. These pumiceous tuffs are clearly detrital and consist largely of albite crystals and fragments of crystals.

Both the Teepookana and Zeehan-road tuffs are interbedded with rocks deposited in water. At Teepookana these are dark grey shales with pyrite (see Table 11, No. 4) which is present as disc-like bodies with crenulate, nibbled margins. In one example, the margins in section appear to have retreated irregularly from a previously perfect oval shape. The bodies lie parallel to the bedding and might very well be of organic origin. The pyritic shale indicates reducing conditions in relatively deep water. On the Zeehan road, the tuffs are interbedded with shales and greywackes; Blissett (1962, p. 40) has described a fossil locality in shales a few hundred metres from the tuffs ~~and lavas~~ described here and the fossils are marine sponges of Cambrian age.

Origin of the Teepookana crystal-vitric tuffs

The origin is uncertain though some reasonable deductions can be made. For the crystal-rich tuffs the significant features appear to be as follows:

They consist of up to 50% of a poorly sorted assemblage of crystals of feldspar and quartz arranged at random with a disrupted framework and not showing flow effects. They possess volcanic rock fragments and up to 10% pumice fragments, again usually in random orientation, and the pumice fragments are usually prominently banded and in some cases contorted and frayed.

The tuffs generally have only the faintest internal stratification and therefore do not appear to be ash showers. Most of them are interbedded with sediments deposited in water and probably were also deposited in water. Though some of the beds vary considerably in grain size and are poorly sorted no definite evidence of grading was noted. They occur w th pale and dark grey "glassy" tuffs at Teepookana and these tuffs differ in having no crystal fragments and in the presence of slivers and contorted lenses of streaky, dark grey, devitrified glass (?). Streaks and lenses of dark glass are characteristic of many ignimbrites eg., the "wilsonite" of New Zealand (Marshall, 1935) and streaky tuffs within welded tuffs have been described from the Lake District of England (Oliver, 1954) and Snowdon (Beavon et al., 1961).

There appear to be conflicting criteria for the origin of these tuffs; they have several of the characteristics of non-welded ignimbrites or froth flows (lack of bedding, fragmental texture, fibrous devitrification of ovoid pumice fragments, collapsed pumice fragments, high glass content etc.). Though there is little doubt that they were deposited in water the expected stratification is rare and hence they are unlikely to be normal ash showers.

They are similar to the sub-aqueous pyroclastic flows from Washington described by Fiske (1963), and in particular to his pumice-rich varieties, which are relatively poor in rock fragments. Fiske suggests these rocks were derived from underwater eruptions that initiated density-current flows of volcanic products off the volcanoes' flanks or were derived from slumping off the flanks. Particular features are their occurrence with normal banded ash-fall tuff; their non-welded nature; the poor sorting, disrupted framework and random orientation of the fragments; their lack of stratification; and the presence of grading and small slump structures in the laminated tuffs. He suggested some of the products resulted from aqueous quenching.

Fiske's mechanisms seem at least partly applicable to the Teepookana tuffs. Some tuffs in the Mt. Read Volcanics are probably subaerial and it is likely that subaerial deposition took place not far from the sea. If the vulcanism involved formation of ignimbrites or froth flows then entry of hot gas clouds or gaseous lava flows into the sea is also to be expected. On entering the water these materials might turn to high-density mud-flows (lahars) composed of crystals, glass shards, pumice etc., and capable of travelling considerable distances as discrete units, or they might even persist as hot gas clouds. Such currents could pass from the Mt. Read Arc well out into the deeper parts of the Dundas Trough. Williams (1941, 1957) has reported nuée ardente flows that turned to lahars on entering river water. Rankin (1960) maintains that a nuée ardente would disperse upward on penetrating water because of its density but this seems an oversimplification of the case. Many pumice fragments would float but glass, crystals and rock fragments would combine to form a relatively high density fluid unit in which steam and water would replace hot volcanic gas as the buoyant medium. There seems no reason why the flow should not continue its downward course as long as the gravity gradient persists, and then dump its load on becoming immobile. The final deposit would form in an aqueous environment, be poorly sorted and consist of varying proportions of crystals (mainly quartz and feldspar), glass fragments, pumice and rock fragments (mainly lava but possibly some country rock from the vent walls).

The effects of quenching on the glassy lava is worth considering. The temperature drop from the hot gas environment at about 900° C (Ross and Smith, 1960) to the temperature of the water would be fairly rapid; rather quicker for instance, than the drop applied to a normal vitric ash shower that passes into the atmosphere in a gaseous cloud and then falls through warmed air into water. This sudden change is likely to cause cracking and disruption of the glass fragments, thereby forming large quantities of glass "dust". Material of this type makes a likely source for the microcrystalline, "dusty" aggregate that makes up several of the fine tuff beds at Teepookana. However, the temperature in parts of these subaqueous flows might have been maintained at several hundred degrees Centigrade for a while, sufficient to cause some streaking and streaming out of fragments of plastic glass material during flow.

A second possibility is that slumping of crystal-vitric tuffs off the flanks of volcanoes in the Mt. Read Arc gave rise to density currents capable of travelling considerable distances, the resulting deposits being similar to those derived from lahars.

A third possibility is that the Teepookana tuffs were derived from glowing avalanches or froth flows that erupted under water within the Dundas sedimentary trough. This seems less likely because there is some doubt as to whether such an eruption could take place, and because keratophyric flows are extremely rare away from the Mt. Read Volcanic Arc.

Fine-Grained Stratified Tuffs

Tuffs with pronounced banding occur on Whip Spur and similar but altered rocks probably crop out on Philosophers Ridge at Mt. Lyell. At Whip Spur (31675) the tuffs consist of finely banded aphanitic material containing a few beds of coarse tuff. They underlie the pyroclastic breccia described on p. 96 and overlie the Darwin keratophyre that forms bold outcrops at the head of Whip Spur. The aphanitic tuffs are greenish grey with pale-coloured cherty bands from ½ mm. to 2-3 mm. thick. The bands tend to weather differentially and are very obvious on weathered surfaces. Thin sections reveal a dusty, very fine-grained to microcrystalline aggregate with scattered grains of quartz and albite up to 0.06 mm. diameter. The banding appears to be due to variations in density of the coarser quartz and albite fragments and to variation in density of the dark, "dusty" matrix. These rocks were described by Bradley (1954) as fine-grained greywacke. Certain bands show considerable crumpling apparently of convolute (intraformational) type. The folds are up to several feet across and consist mainly of elongate paraboloids or brachydomes with axial surfaces perpendicular to the bedding but with random strike (Plate 41).

Fine-grained stratified tuffs occur in the Rosebery and Hercules Mines. The cross sections of these mines are similar and indicate black slate overlying "host rock" described as tuffaceous shale (Hall et al., 1953, p. 1150, and Fig. 12). The "black slate" is finely banded and fine grained, the bands varying from dark to pale grey. To obtain a complete cross section of the host rock and black slate, the core from boreholes R1028, R844 and R918 (Rosebery mine, 12 level) were examined, as well as the surface exposures in the open cut (Fig. 59). In most of these rocks, the degree of sericitisation is fairly high and there is considerable recrystallisation of quartz near the sulphides. Feldspar crystals are only just discernible in thin section. However, the fragmental nature and crystal outlines show up well on cut and polished faces and for this reason the diamond drill core is much easier to examine than surface specimens. The combined, summarised log reads as follows, beginning at the hanging wall (note that the thicknesses are

Plate 41 No. 1:- Paraboloidal folds in banded tuffs, Whip Spur, looking down on bedding surface.

Plate 41 No. 2:- Paraboloidal folds in banded tuffs, Whip Spur, looking obliquely along bedding surface.

as measured in the drill hole and are not stratigraphic thicknesses):

6.5m: Massive Pyroclastics - creamy or pale green sheared, streaky rock (31777) consisting of a very heterogeneous fragmental aggregate of quartz and very altered feldspar crystals.

20m: Irregular alternations of dark and pale grey, finely laminated tuff or tuffaceous shale with lenses of coarse feldspathic tuff (31776-31774). The fine tuff consists of a heterogeneous, very fine-grained quartz-sericite aggregate. The coarse tuff has a similar matrix in which are scattered at random embayed quartz grains, albite and a few rock fragments. The laminations of the fine tuffs are in many cases tightly folded about the cleavage but in much of the rock the bedding has been destroyed by the cleavage. These beds contain conformable lenses of coarse-grained calcite that is pre-cleavage and presumably syngenetic. Though no obvious sedimentary material was encountered in the borehole, some of the rocks in the open cut (e.g. 31856) are fine to coarse-grained siltstones with a fine sericitic, opaque matrix and quartz and feldspar grains up to 0.02mm. diameter. The siltstones are cut by quartz veinlets up to 0.3mm. thick that are sub-parallel to the bedding, and closely spaced, and contain ragged crystals of pyrite.

These veinlets (and the pyrite ?) are boudinaged and disrupted by the almost parallel foliation and are therefore pre-cleavage (Plate 42, No. 1).

13 m.: Medium grey, medium-grained feldspathic tuffs with thin 1-2 cm. bands of dark grey tuff and some layers of coarse-grained tuff (31773, 31781).

18 m.: Pale greenish-grey, fine-grained sericitic tuff with streaks and bands of dark grey material. Some crumpling locally but cleavage has obliterated most of the bedding (31769-31662). Some layers with conglomeratic appearance.

Orebody: Conformable lens of massive sulphides or sulphide-rich quartz sericite "schist".

94 m.: Streaky, fine-grained feldspathic (?) tuffs with patches of very fine-grained schistose sericitic rock (32913-32917). From 270 to 300 cm. the schistose rocks are cut by a dark fine-grained pyroxene-labradorite basalt dyke (32912) that is post-mineralization and post-cleavage and may be of Tertiary or Mesozoic age. Similar basalt dykes cut the Renison Bell sulphide orebodies. The coarser tuffs (e.g. 32917) contain quartz crystals and sericitic lath outlines in a very fine-grained sericitic base, partly replaced by carbonates; the texture is distinctly fragmental. The more

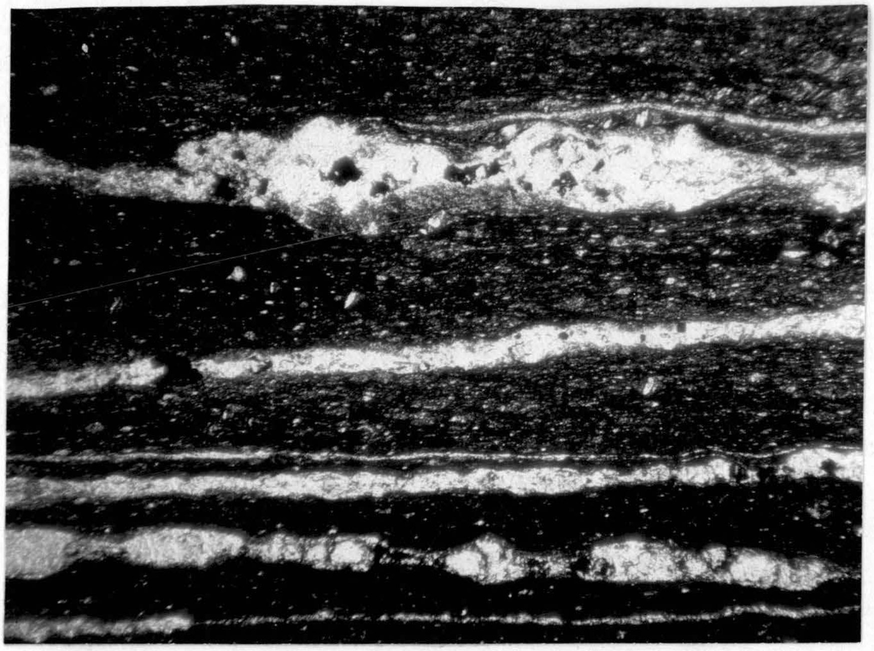


Plate 42 No. 1:- Boudinaged quartz-veinlets in slates at the Rosebery Mine. X20.

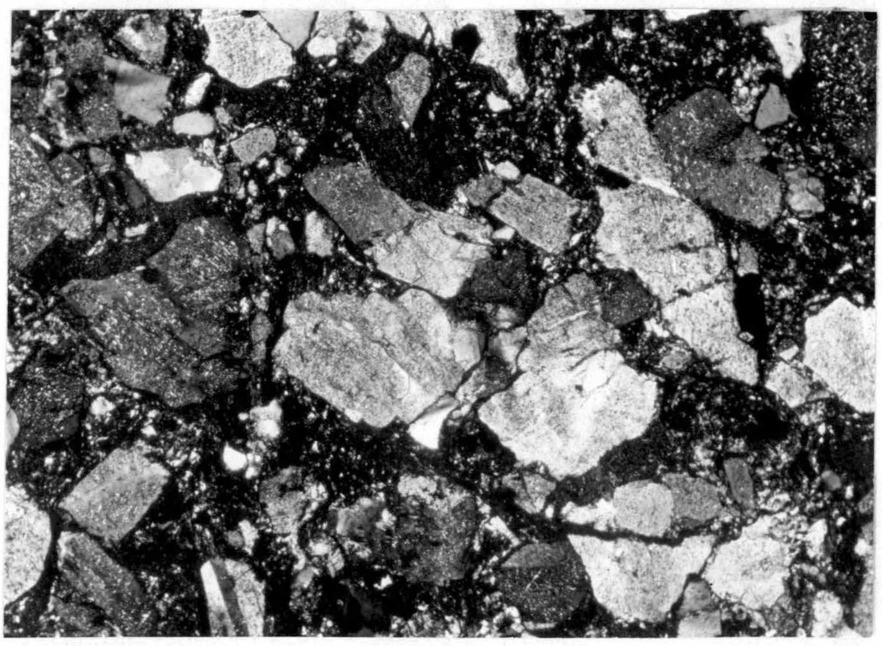


Plate 42 No. 2:- Crystal tuff, Lynch Creek, crossed nicols. X25.

finely grained rocks consist of schistose, almost microcrystalline, quartz-sericite aggregates (e.g. 32914).

30 m.: Pale grey sericitic quartz schist with fine-grained quartzose matrix (31778, 31779, 31784A, 31785A, 31786A, 31774A). This passes down to sheared quartz-feldspar tuffs and lavas of the Primrose Volcanics that have been described elsewhere (31791, 31789A etc.).

It appears that the "host rocks" of Rosebery are mainly tuffs and that though much of the "black slate" is similar, at least some of it is of sedimentary origin. The well-developed stratification indicates deposition in an aqueous environment. Possibly the Rosebery and Hercules banded tuffs and siltstones were deposited in small lakes of ephemeral nature within the growing volcanic chain along the West Coast Range.

Beds of dark and pale grey laminated tuffs, siltstones and fine greywackes occur in many parts of the Mt. Read Volcanic Arc, the proportion of volcanic and sedimentary material varying considerably. In South Owen Creek they form a unit up to 100 m. thick and extending for nearly a mile along strike. The banding of rock types within this unit is impersistent and variable. The bulk of the unit (31629, 31631, 32549) is made up of siltstone composed of a fine-grained quartzose aggregate with sericite and occasional larger grains of quartz and albite up to 0.03 mm. diameter. A bulk chemical analysis of the unit is given in Table 11 (No. 11).

The sediments forming a four or five mile belt from north of Tullah to the Stirling mine below Mt. Murchison (the Farrell slates - Fig. 23) are very similar, (30013, 14, 15) but contain more thin beds of tuff. These consist of altered albite laths and broken quartz crystals up to 0.75 mm. diameter in a very fine, "dusty" matrix.

Finely stratified tuffaceous rocks and siltstones are interbedded with quartz keratophyre on Roaring Meg Creek (32844, 31230, 31231, 31232, 31672) and very delicately banded tuffs are common between the Henty and Yolande Rivers on the Queenstown-Zeehan road.

In all these tuff-siltstone units thin (3-½ cm.) bands of contorted or "balled" tuff are common within the fine-grained bands; the deformation is probably related to the period of sedimentation. Rock fragments are rare and consist of tuff and fine-grained lava (e.g. 31672). They are characterised by pronounced stratification in which individual bands are not persistent; in most areas the units are at least several tens of feet thick and they persist along strike for distances varying from a half to several miles. The sorting implied by the stratification implies deposition in water, probably from ash showers, and an aqueous environment is also indicated by the intraformational crumpling.

Crystal Tuffs

Coarse crystal tuffs occur as bands and lenses within the fine-grained stratified tuffs and siltstones just described, varying from a few cm. to several metres thick, and tuffs are also interbedded with lavas and with sediments. A considerable variety of these tuffs occur south-east of Queenstown, in the Roaring Meg Creek and Lynch Creek areas. For example, in Lynch Creek a siltstone unit which overlies several hundred feet of spilites and includes spilitic flows, contains a 3 m. bed of quartz-feldspar tuff (31657); this consists of a chaotic aggregate of albite and quartz crystals in a sparse matrix (Plate 42, No. 2). An analysis of this rock is given in Table 11 (No. 9). As can be seen from Table 32 it contains remarkably high values for copper and zinc. In places the rock contains lenses and slabs of crumpled, banded tuffs and blocks of feldspar porphyry up to several feet in length. Similar tuffs occur south of Tullah (31782) in the Farrell slates.

32867 from Roaring Meg Creek is unusual in that it contains several percent of rock fragments (mainly lava) and fragments up to 1/3 mm. diameter of microcrystalline aggregate (devitrified glass) showing perlitic cracks (Plate 43, No. 1).

Quartz-rich tuffs associated with spilites also occur at Zeehan and at Curtin Davis on the N.E. Dundas tramway. At the latter locality (50G135, 146-151, 154-158, 169, 172) the rocks consist mainly of albite laths (both fresh and sericitised) quartz shards and fragments, and a few fine-grained, tuff-like, rock fragments in a dusty matrix. Proportions of quartz and feldspar vary and maximum fragment size is 4 mm. diameter. Similar tuffs are associated with potash rhyolites and breccias, e.g., at Lake Jukes and Jukes Pty. (32862), and at Rosebery and Hercules in the Primrose Pyroclastics and the Mt. Read Volcanics (606, 607, 603). Tuffs without quartz are in many cases difficult to distinguish from lavas though they usually show fragmental texture (e.g. 32550, 32880, 32860, from the Great Lyell-Lynch Creek area).

Above the Rosebery "slate" bed are volcanic breccias and crudely banded tuffs (31796A), the Massive Pyroclastics of Hall et al. (1953). They are particularly well exposed on the Tullah road at the Rosebery town limit. They are essentially crystal tuffs with albite crystals up to 2 mm. diameter scattered randomly in a medium-to fine-grained (recrystallised ?) feldspar-quartz-

Plate 43 No. 1:- Perlitic cracks in fragments of glass, Roaring Meg Creek. X45.

Plate 43 No. 2:- Crude banding in keratophyric tuff, Lynch Creek.

sericite matrix. Irregular and impersistent ~~but~~ subparallel bands of aphanitic pale grey rock cut the tuff and are up to 1 cm. thick; they appear to be merely more finely grained versions of the tuff. Some of this fine material appears to be intrusive because dyke-like sheets up to 6 cm. thick cut steeply across the banding described above. Most of these crystal tuffs were deposited in water, some as ash showers, others as pyroclastic flows carried into the aqueous basins by density currents.

Keratophyric Tuffs with Ferromagnesian Minerals

Tuffs containing ferromagnesian minerals are rare. An augite-albite tuff from the Lynch Creek area has been described previously by the writer though it was suggested then that it might be a lava. Bradley (1954) suggested it was a metasomatised conglomerate. It forms a bed up to about 300 m. thick extending along strike for at least three miles (32908, 31659). It consists largely of stumpy albite crystals that are remarkably fresh. These may be scattered throughout a microcrystalline dusty aggregate or be so closely packed that very little matrix is present. Up to 30% of the rock consists of fresh diopsidic augite crystals and there are patches rich in rock fragments, mainly fine-grained basalt, keratophyre and hornblende keratophyre. These rock fragments are up to 60 x 20 cm. in size and locally give the rock the appearance of a breccia. Quartz forms up to 10% of the rock and chlorite replaces the matrix locally (β =1.615). The fragmental appearance of the rock is sufficient to indicate its pyroclastic origin and this is further supported by a crude banding seen in some outcrops (Plate 43, No. 2). It appears to underlie the Lynch Creek spilites and probably represents the initial phases of spilitic eruption. Its composition is akin to a keratophyre (Table 11, No. 10). An identical rock occurs on Little Owen ridge and locally appears to cross-cut the surrounding rocks; it may in part occupy a local volcanic vent (Solomon, 1957). The freshness of the albite in this and other feldspathic tuffs is of some interest, particularly at Lynch Creek where the feldspars in the overlying spilites are extremely altered, in some cases consisting of coarse aggregates of micas. A possible reason is that the tuffs are not pervaded by late stage gases or solutions to the same extent as lavas, the tuffs therefore not representing the complete composition of the extruding magma.

On the slopes north of the upper zig-zag on the Comstock tram line there are outcrops of a quartz-feldspar rock showing crude coarse banding (32530). It is a fragmental rock, composed of about 65% albite laths and 20% quartz crystals and fragments whichaverage 1 mm. grain size. The dusty matrix varies in colour, being darkened by iron ore dust and made green by chlorite. The crude banding is due to chlorite enrichment in layers. The chlorite is most probably of primary or diagenetic origin.

Depositional Environment of the Keratophyric Tuffs

Over most of the West Coast, away from the Mt. Read Volcanic Arc, keratophyric tuffs are associated with non-terrestrial sediments and/or sediments with marine fossils, e.g. Teepookana, Henty River, Lynch Creek, Zeehan etc. Some of these tuffs are ash falls or pyroclastic flows and most were probably deposited in water, though the source volcanoes may have projected above the water surface. Some of the Teepookana tuffs may have originated as subaerial gaseous flows that swept into water or have formed by slumping from volcanic deposits.

The role of aqueous deposition in the Mt. Read Volcanic Arc is not clear. The Natone Volcanics are interbedded with aqueous sediments, and finely stratified tuffs and siltstones presumably of aqueous origin occur in the lower Mt. Read Volcanics

(Primrose Volcanics) near Rosebery and elsewhere. Combining this information with that given by the lavas, it appears that part of the basal portion of the Mt. Read succession is aqueous and part probably subaerial. The lack of finely banded rocks in the upper part of the Mt. Read Volcanics suggests that more and more of the Arc became emergent in the later stages of vulcanism.

GENESIS OF THE FELDSPARS IN THE VOLCANICS

Albite

The almost ubiquitous albite has similar optical properties in all rock types (including gabbros, granites and fragmental rocks) and, in common with albite in other spilite-keratophyre assemblages, has somewhat anomalous properties. Though the albites are clearly low-temperature in type, twin axis measurements show considerable departures from ideal curves and the 2V readings are between the high and low temperature curves. Similar features have been noted by Donnelly (1963) on Virgin Island rocks and by Valiance (1960) in New South Wales rocks. Donnelly referred to crystals which have high 2V values (90-100) as possessing quasi-low temperature (QLT) optics. Experiments by Mackenzie (1956) indicate that disordered albite held at a few hundred degrees

Centigrade for several years would become ordered. Therefore QLT optics should not survive either slow cooling or later regional metamorphism. Donnelly (1963) used this argument to prove that his anomalous albites are volcanic and not metamorphic. The Tasmanian volcanics must have undergone mild burial metamorphism as, at least in some cases, they have been covered by up to ten thousand feet of younger sediments. Similar remarks apply to the New South Wales rocks. Therefore the tendency to QLT optics can hardly be taken as proof of volcanic origin. For mild departures from the ordered state, factors such as substitutions, superlattice developments etc., may be the cause of anomalous optics. Vallance (1960, p. 39) has suggested that LT optics in albite may not be "specially critical", in view of the rarity of HT albite in lavas general y.

Donnelly suggested that curved phenocrysts may be significant in determining albite genesis. He suggested that the curvature would not be preserved during replacement of a calcic feldspar by albite and that curved crystals must have been albite (and possibly ordered albite) before bending. Though not severely bent, the few curved phenocrysts observed in the Tasmanian rocks are not recrystallised and the groundmass is not deformed (curved crystals in cataclasites are easily distinguished by the shearing of the matrix). Therefore, the crystals were bent before solidification of the rock. It does not seem possible to continue the argument beyond this, as Donnelly has done, because possibly the bent crystals could have been, say, andesine, that was replaced by albite after bending.

The presence of andesine in the hornblende andesite suggests that a parental calcic feldspar may have been replaced by albite. Although no stage in this process has been observed, a "secondary" origin for the albite (i.e. derived from earlier crystals of different composition) must remain a possibility.

The presence of ophitic texture between albite and augite indicates primary albite but again replacement could have taken place. The albite phenocrysts in many rocks are much altered and the influence of deuteric solutions has obviously been considerable in developing the mineralogy of these rock. They are also known to have been deeply buried and folded, and there is therefore the possibility of a non-volcanic origin. However, a primary volcanic origin seems likely for most of the albite.

Potash Feldspar

The K-feldspars offer similar genetic problems to those of the albites. In the spherulitic types the K-feldspar may be the result of devitrification or it may represent primary crystallisation;

/

either way there is no evidence of replacement of an earlier mineral. In the non-spherulitic types, some of the K-feldspar crystals are crudely developed and might be secondary replacements but in others the crystals are moderately well developed and give no evidence of replacement.

K-feldspar rims on albite cores in the pumiceous flows are evidence that the K-feldspar crystallised later than the plagioclase, a feature commonly found in acid igneous rocks. The pumice surrounding the K-feldspar rims in rocks 32956, 50G162, etc. suggests that the rims developed prior to consolidation of the flows. Alkaline silicates are highly soluble in gaseous water (Morey and Hesselgesser, 1951) and thus crystallise late and are easily distributed throughout such permeable materials as semi-consolidated froth flows. The abundance of K-feldspar in the groundmass of several of the lavas (e.g. at Darwin and Mt. Sedgwick), also indicates late crystallisation for the K-feldspars.

However, a study of the compositions of these rocks suggests a non-magmatic origin. The analyses have $>$ 80% normative Ab + Or + Q and therefore are close to petrogeny's residual system, which has been investigated experimentally (Bowen, 1937 and later papers). Residual compositions derived from a melt of basaltic composition should fall in the thermal

Fig. 14: Diagram representing the $NaAlSi_3O_8$-SiO_2 system

(a) Showing (1) Bowen's thermal valley (Bowen, 1937) (2) isobaric field boundaries for water vapour pressures of 500 and 2000 atmospheres (using the diagram as the anhydrous base of the $NaAlSi_3O_8$-$KAlSi_3O_8$-SiO_2-H_2O tetrahedron). (3) field boundary AB proposed by Steiner (1963) as compared to Bowen's boundary AA'.

(b) Showing distribution of normative albite, orthoclase and quartz in rocks carrying 80% or more normative (Ab + Or + Q). Dashed lines for 571 plutonic rocks, full lines for 362 extrusive rocks, at concentrations of 1% and 5%, after Tuttle and Bowen (1958, p. 78-79).

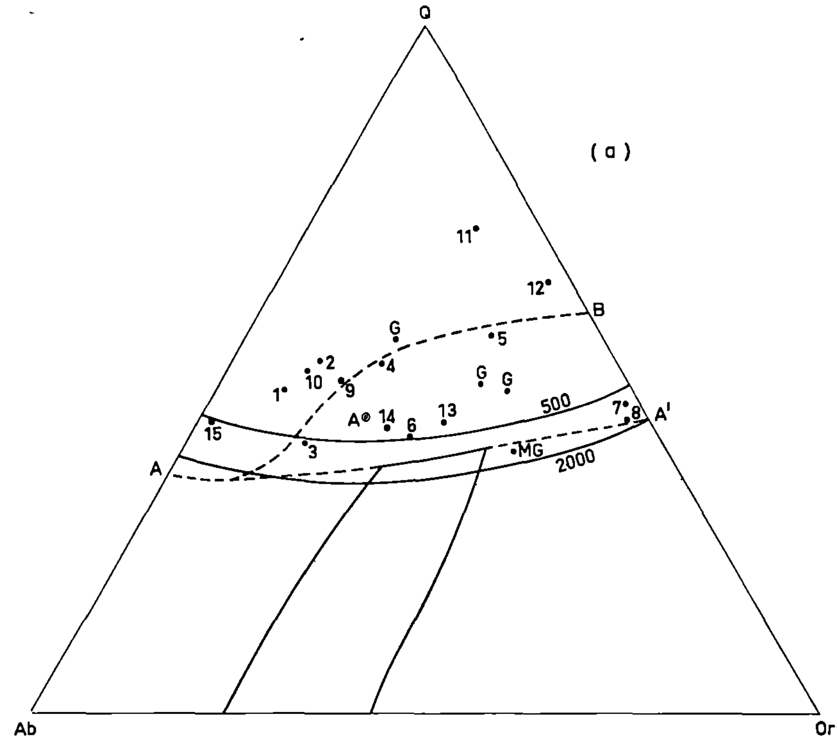
Tasmanian Cambrian Rocks with >80% (Ab+Or+Q):-

1 Quartz keratophyre of Great Lyell (31702).
2 Quartz keratophyre in Waterfall Gully (31272).
3 Quartz keratophyre in West Queen River.
4 Quartz keratophyre in East Queen River.
5 Quartz keratophyre, Yolande River Bridge (31717A).
6 Darwin keratophyre, Whip Spur (31676).
7 Darwin keratophyre, Intercolonial Spur.
8 Darwin keratophyre, Mt. Sedgwick (31257).
9 Quartz keratophyre, East Queen River (31754).
10 Primrose Volcanics, near Rosebery Mine (R265).
11 Primrose Volcanics, Williamsford Road (R176).
12 Primrose Volcanics, near Barkers Crossing, Rosebery (R1).
13 Primrose Volcanics, Rosebery (R177).
14 Primrose Volcanics, north of Hercules (31789A).
15 Pale cherty tuff (3177A) west of Teepookana, King River.

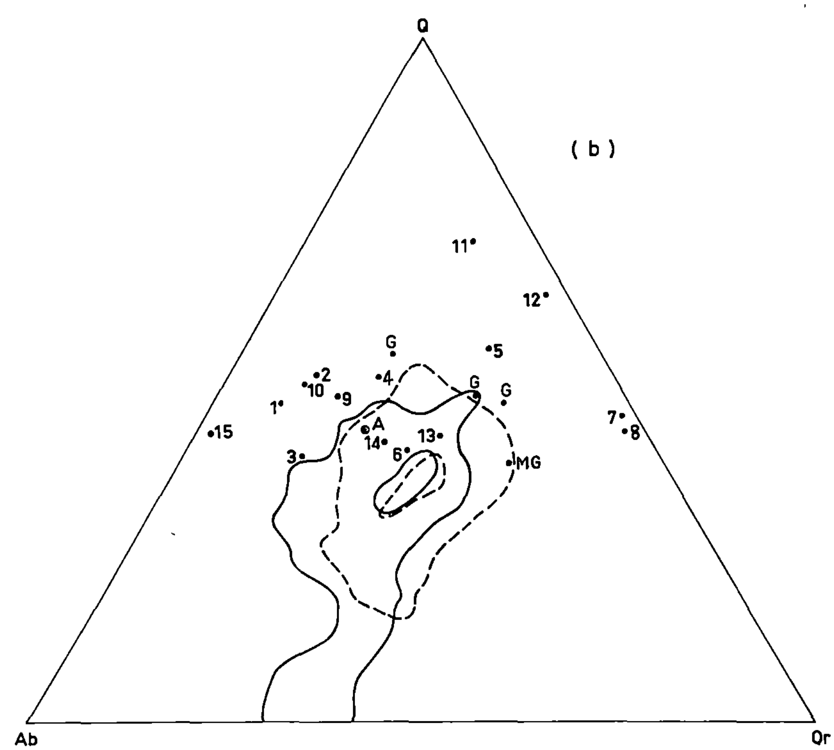
A Average of volcanics 1 to 15.

G Darwin Granite.

MG Murchison Granite.



Q
(a)
11
12
B
G
5
2
4
G
G
10
1
9
500
7
8
A'
A
14
6
13
15
3
MG
2000
Ab
Or



Q
(b)
11
12
G
5
2
4
G
G
10
1
9
A
7
8
15
14
6
13
3
MG
Ab
Or

along the wet trough but there is nevertheless a wide spread over the diagram. Analyses 11 and 12 ~~are~~ quite possibly represent rocks silicified during mineralization. Even allowing for non-isobaric crystallisation (a highly probable condition) it is difficult to explain the analytical results in terms of magmatic differentiation. However, many of the problems disappear if it is assumed that potassic and sodic magmatic fractions can separate before extrusion from a sodi-potassic magma. The average (A) of all the quartz keratophyres plotted on Figure 14 falls comfortably within the residual thermal valley and the objections to a magmatic origin with regard to composition, at least as far as the alkalies are concerned, may not be quite so condemnatory as at first appears.

Boone (1962) has discussed the separation of a potassic-rich syenite melt from a granitic magma. In a magma chamber there would tend to be a concentration of water vapour near to the chilled margins and the roof because of the temperature gradient (Kennedy, 1955, p. 491). This upward transfer of water would tend to cause a similar concentration of the alkali silicates which are particularly soluble in supercritical water (Morey and Hesselgesser, 1951). Boone (1962, p. 147) suggested that the transfer of potassium ions "should be more effective" than other ions because of their relatively large specific volume. On release of pressure (eruption) the potassium ions form K-feldspar which may develop as crystals or as rims upon a nucleus of earlier precipitated albite in equilibrium conditions, or may commence to replace albite if the composition of the residual fluid has been heavily enriched in potassium ions.

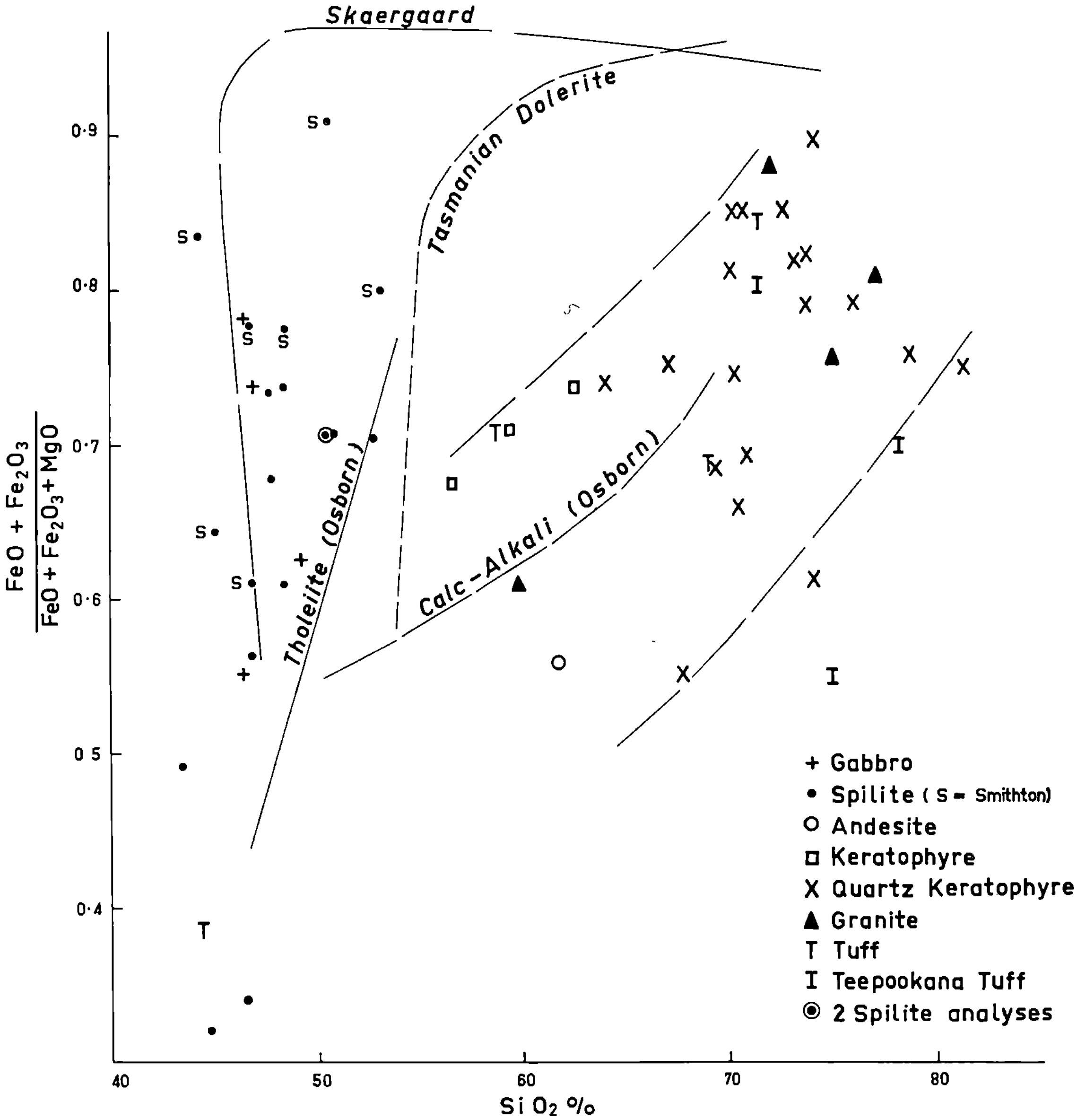
Initial phases of eruption, then, are likely to be vapour rich and potassic while later eruptions are likely to be sodic. This, of course, is precisely the trend in the Rosebery area.

CHEMICAL COMPOSITION OF THE KERATOPHYRIC LAVAS AND FRAGMENTAL ROCKS

The essential features of both the intermediate and acid keratophyric lavas have already been mentioned and comparison with average figures is given in Tables 8 and 9.

Plotting these analyses, and those from Tables 11 and 12, on the variation diagrams used for the spilites indicates strong calc-alkaline tendencies. Note that all of Finucane's analyses for the Primrose Volcanics have been shown as quartz keratophyres though many of them may well be pyroclastic. The SiO_2 v mafic index diagram (Fig. 15) shows a calc-alkaline trend though on a very broad band, and so also does the SiO_2 v Iron Oxide diagram

Fig. 15: Tasmanian spilites, keratophyres and associated rocks plotted on a mafic index v SiO_2 diagram. Also shown are the differentiation trends for Skaergaard, Tholeiitic basalt, Tasmanian dolerite, and calc-alkaline magmas.
In addition to the spilites of Table 5 there are shown an average for a Type 1 pillow and the analysis of a part of Type 2 pillow (Table 2). These same analyses are shown on Figs. 16-20.



Skaergaard
Tasmanian Dolerite
Tholeiite (Osborn)
Calc-Alkali (Osborn)
0·9
0·8
0·7
0·6
0·5
0·4
40
50
60
70
80
Si O2 %
FeO + Fe2O3 / FeO + Fe2O3 + MgO
+ Gabbro
• Spilite (S = Smithton)
O Andesite
□ Keratophyre
X Quartz Keratophyre
▲ Granite
T Tuff
I Teepookana Tuff
◉ 2 Spilite analyses

(Fig. 18). On the mafic v felsic index diagram (Fig. 16) which emphasises the difference between tholeiitic and calc-alkaline trends, the quartz keratophyres show a fairly regular path similar to that of the Scottish Caledonian igneous rocks (as given by Nockolds and Mitchell, 1948, Table 1) and the Garabal Hill-Glen Fyne complex (Nockolds, 1941), both regarded as calc-alkaline (Nockolds and Allen, 1953). Plotting $CaO/(CaO + \text{alkalies})$ v SiO_2 (Fig. 17) reveals a calc-alkaline trend but in the $Na_2O/(Na_2O + K_2O)$ diagram the spread of analyses is wide. The alkali v SiO_2 (Fig. 18) trends show calc-alkaline tendencies and Rittmann's index is within his Pacific (calc-alkaline) band. The plots of the oxides v SiO_2 (Fig. 18) have calc-alkaline tendencies though the individual alkali diagrams, of course, show a wide scatter at the acid end. This scatter is better demonstrated by the K - Na-Ca diagram (Fig. 19a) and compares closely with other spilite-keratophyre provinces (e.g. Battey, 1956). The often-used AFM diagram (Fig. 19) shows a characteristic basaltic differentiation trend towards the alkali corner but with a wide scatter of analyses. Scattering on variation diagrams is typical of both spilites and keratophyres and is probably related to deuteric alteration and to fractionation during an unusual course of differentiation.

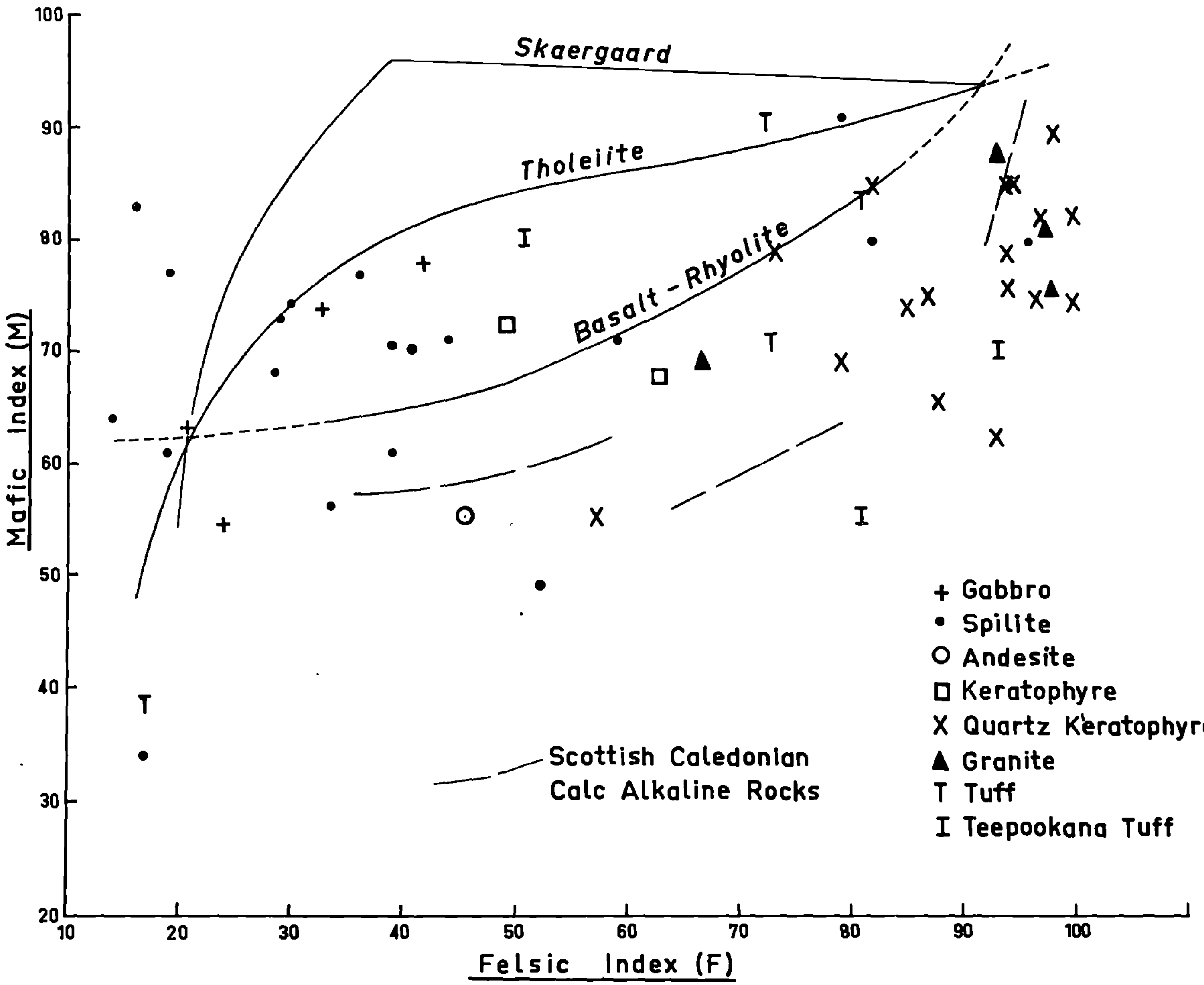
Fig. 16: Tasmanian spilites, keratophyres and related rocks plotted on a

Mafic Index $\frac{FeO + Fe_2O_3}{FeO + Fe_2O_3 + MgO}$

v Felsic Index $\frac{Na_2O + K_2O}{Na_2O + K_2O + CaO}$

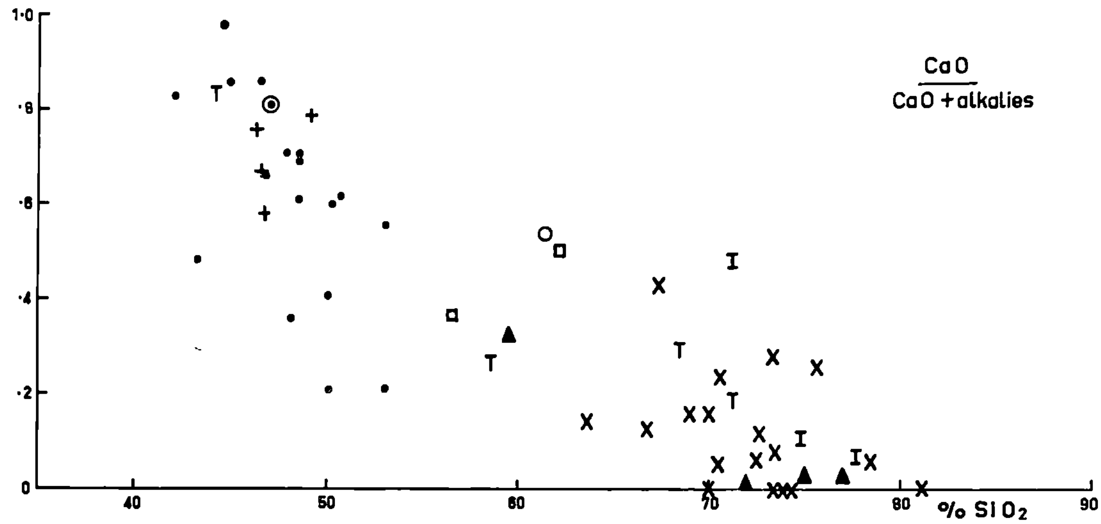
diagram. Also shown are the differentiation trends for Skaergaard, Tholeiitic basalt and calc-alkaline basalt magmas (after Simpson, 1954). Long dashed lines link analyses of Scottish Old Red Sandstone Plutonic, Hypabyssal and volcanic rocks given in Nockolds and Mitchell (1948, p. 534).

Legend as in Fig. 15.



Skaergaard
Tholeiite
Basalt - Rhyolite
Mafic Index (M)
Felsic Index (F)
100
90
80
70
60
50
40
30
20
10
20
30
40
50
60
70
80
90
100
Scottish Caledonian Calc Alkaline Rocks
+ Gabbro
• Spilite
O Andesite
□ Keratophyre
X Quartz Keratophyre
▲ Granite
T Tuff
I Teepookana Tuff

Fig. 17: Tasmanian spilites, keratophyres and related rocks plotted on diagrams of $\frac{CaO}{CaO + \text{alkalies}}$ and $\frac{Na_2O}{NaO_2 + K_2O}$ against silica. Legend as for Fig. 15.


CaO
CaO + alkalies
1·0
·8
·6
·4
·2
0
40
50
60
70
80
90
% SiO2

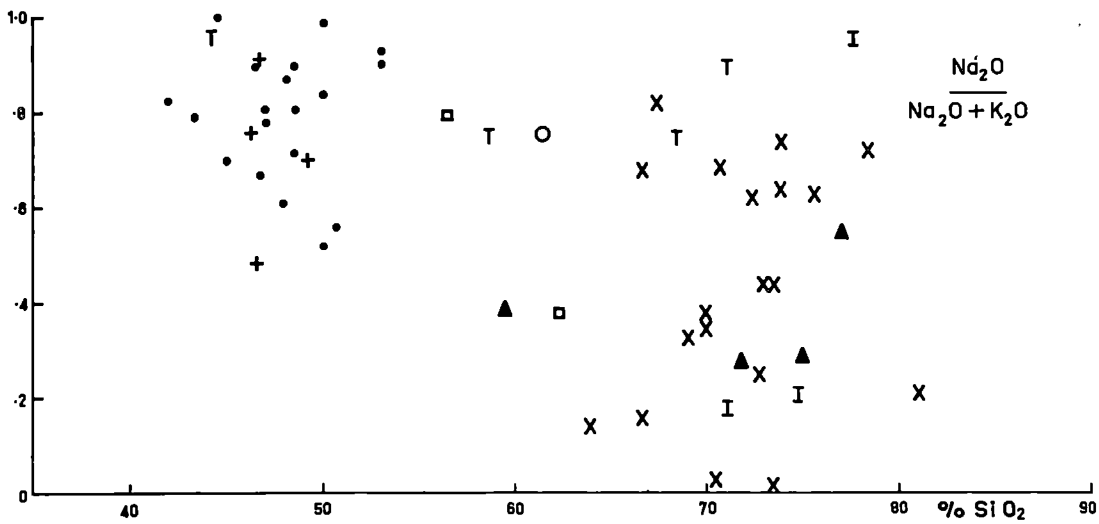
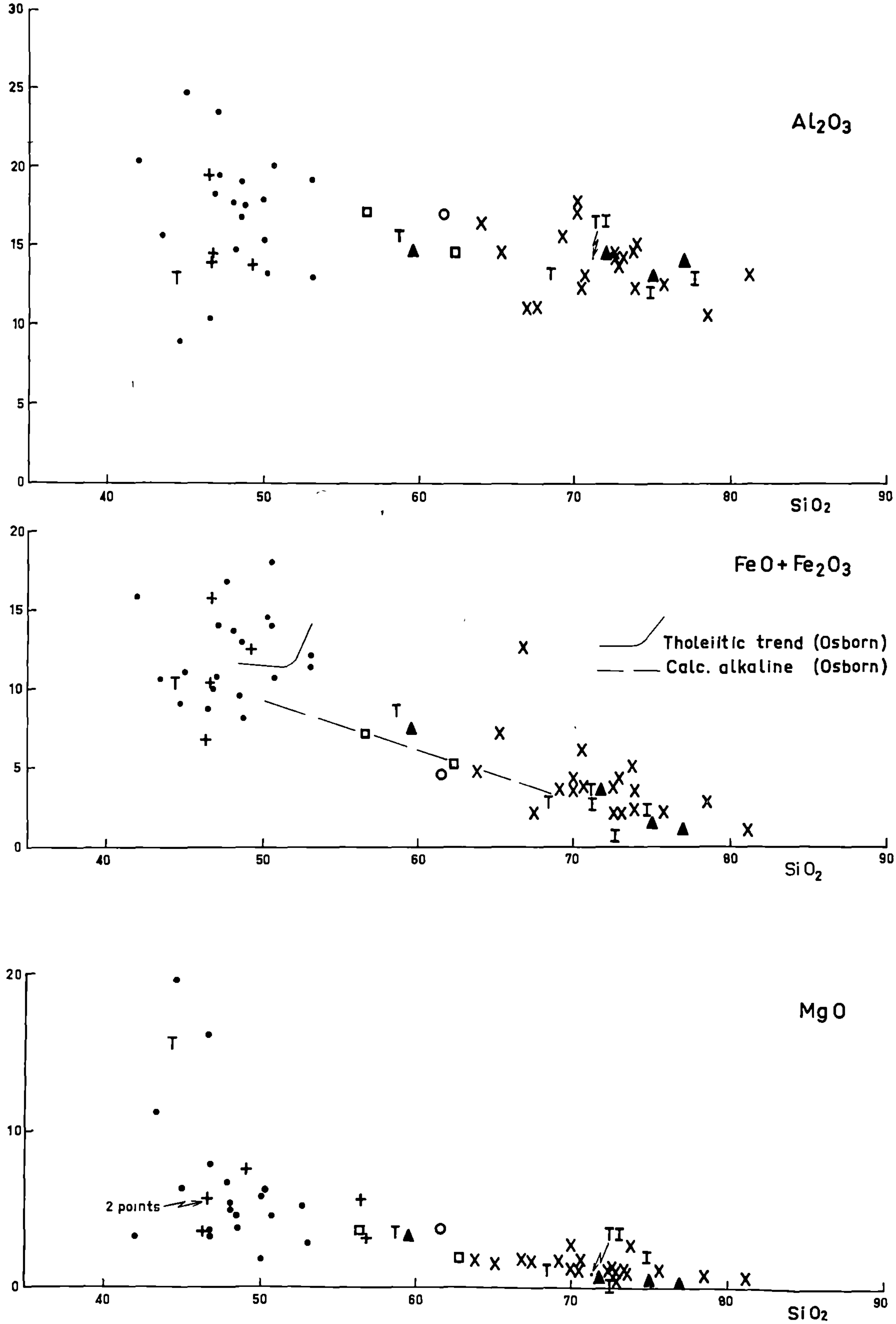
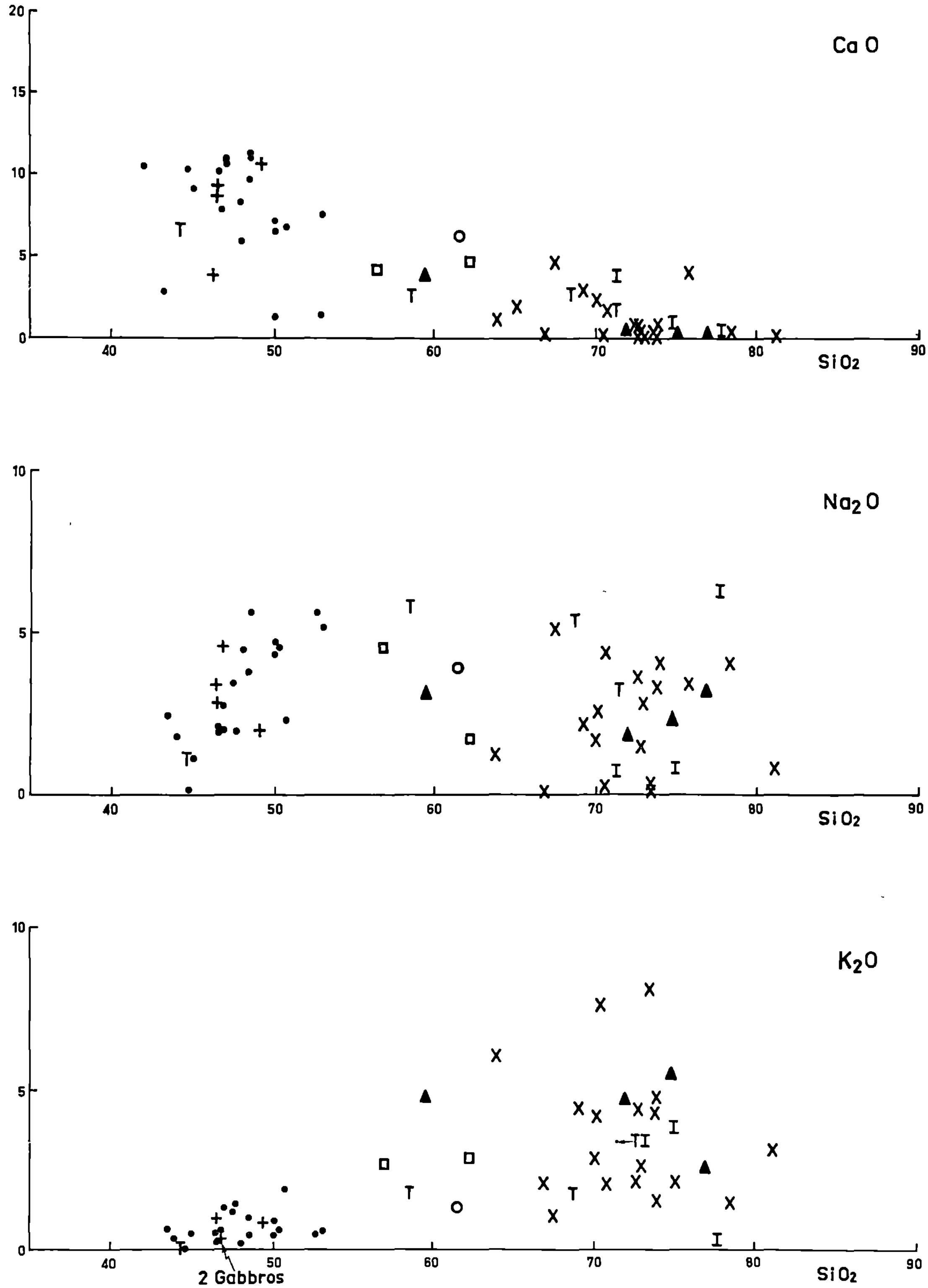

Na2O
Na2O + K2O
1·0
·8
·6
·4
·2
0
40
50
60
70
80
90
% SiO2

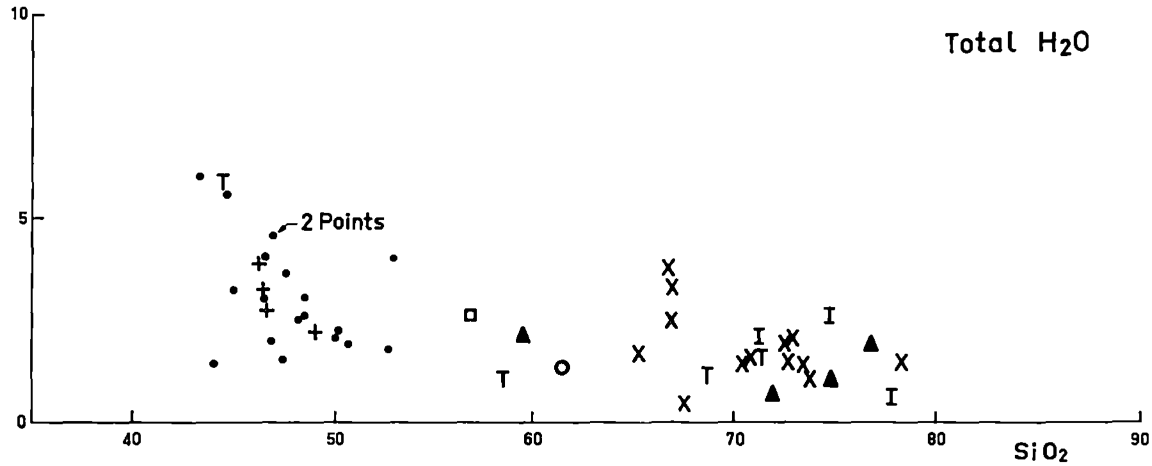
Fig. 18: Tasmanian spilites, keratophyres and related
(3 pages) rocks plotted on diagrams of oxide and oxide combinations against silica. Legend as for Fig. 15.



Al_2O_3
SiO_2
$FeO + Fe_2O_3$
Tholeiitic trend (Osborn)
Calc. alkaline (Osborn)
SiO_2
MgO
2 points
SiO_2



CaO
Na2O
K2O
SiO2
2 Gabbros


Total H_2O
2 Points
10
5
0
40
50
60
70
80
90
SiO_2

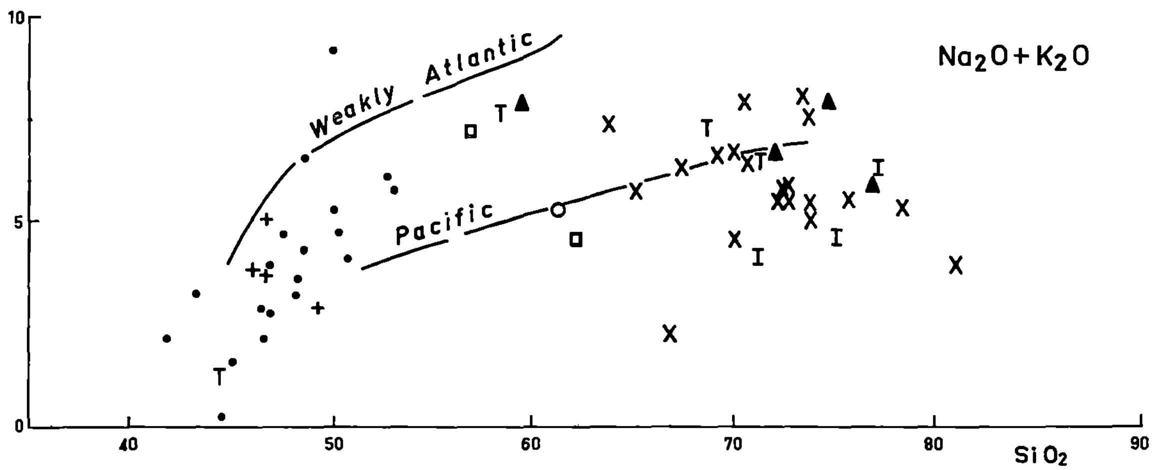
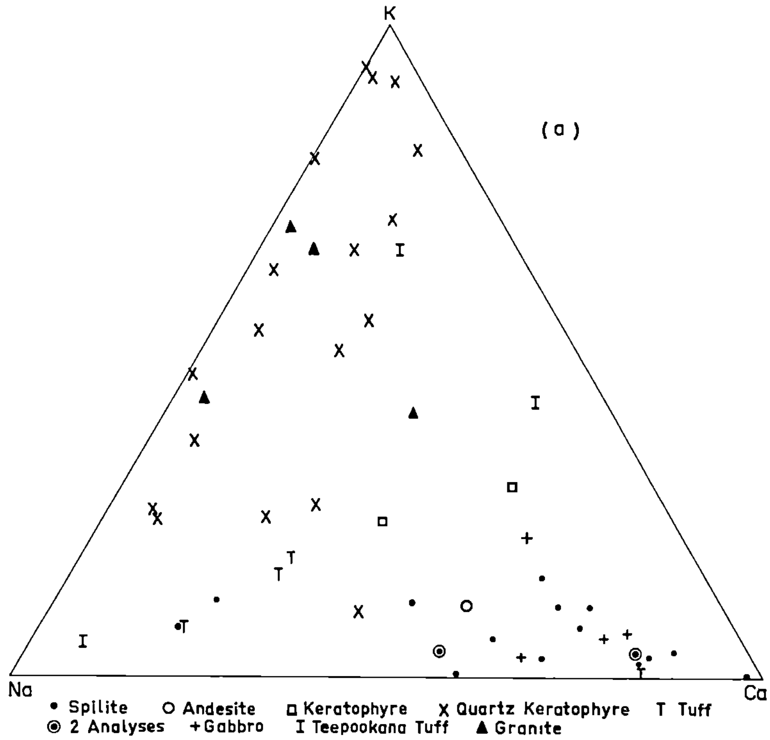

Na_2O+K_2O
Weakly Atlantic
Pacific
10
5
0
40
50
60
70
80
90
SiO_2

Fig. 19: Tasmanian spilites, keratophyres and related rocks, plotted on a K-Na-Ca diagram, and on on Fe-Mg-Alkali diagram.


K
(a)
Na
Ca
• Spilite O Andesite □ Keratophyre X Quartz Keratophyre T Tuff
⊚ 2 Analyses + Gabbro I Teepookana Tuff ▲ Granite

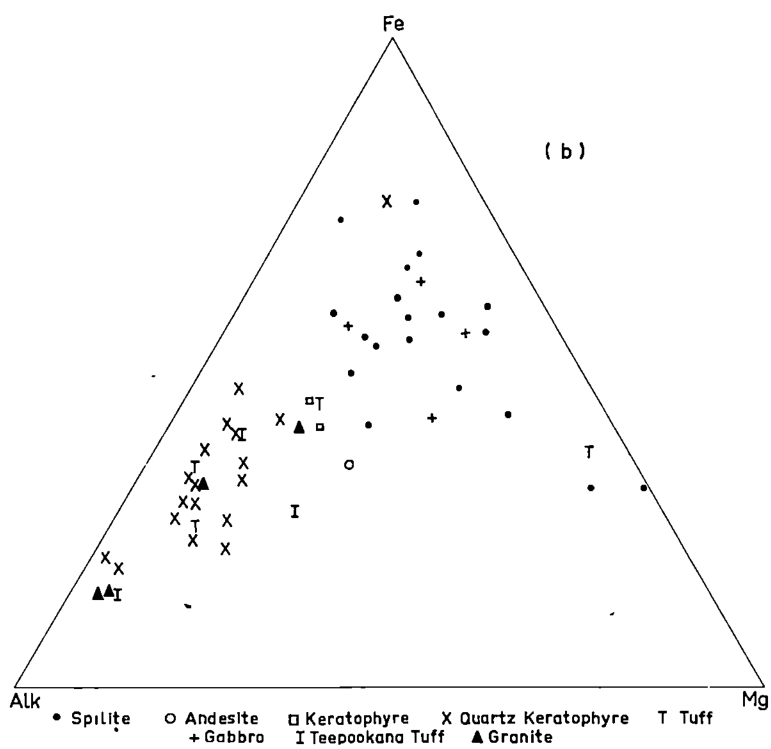
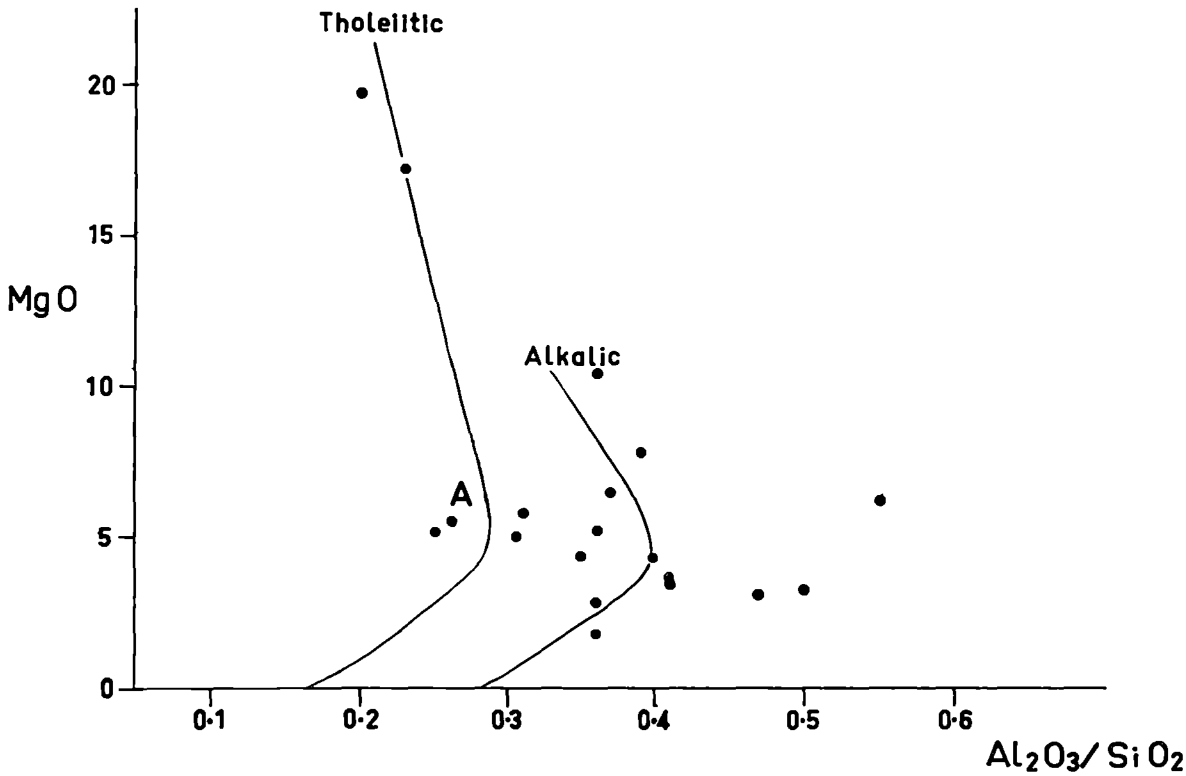

Fe
(b)
Alk
Mg
• Spilite O Andesite □ Keratophyre X Quartz Keratophyre T Tuff
+ Gabbro I Teepookana Tuff ▲ Granite

Fig. 20: Tasmanian spilite analyses (as in Fig. 15) plotted on an MgO v Al_2O_3/SiO_2 diagram (after Eaton and Murata, 1962).



Tholeiitic
Alkalic
A
MgO
20
15
10
5
0
0·1
0·2
0·3
0·4
0·5
0·6
Al_2O_3/ SiO_2

Trace Elements in the Keratophyric Rocks

Available trace element determinations are shown in Table 13. Nickel seems unusually high throughout this spilite-keratophyre assemblage but the figures for the keratophyres may not be reliable. Both copper and lead and zinc are high relative to other calc-alkaline suites, a feature referred to later in discussing the origin of the West Coast ore deposits. Variations of trace elements with silica contents for spilites and more acid rocks are given in Fig. 21 for Ba, Sr, Ni, Rb, Cr and V, these being the only ones showing significant trends. In several cases there appears to be a distinct break between the spilitic and keratophyric rocks, a feature already remarked upon with reference to the major components. The only other trace element determinations on keratophyres known to the writer are by Amstutz (1953, 1954) and his results appear to be within the range of those of "normal" igneous rock series.

Table 13

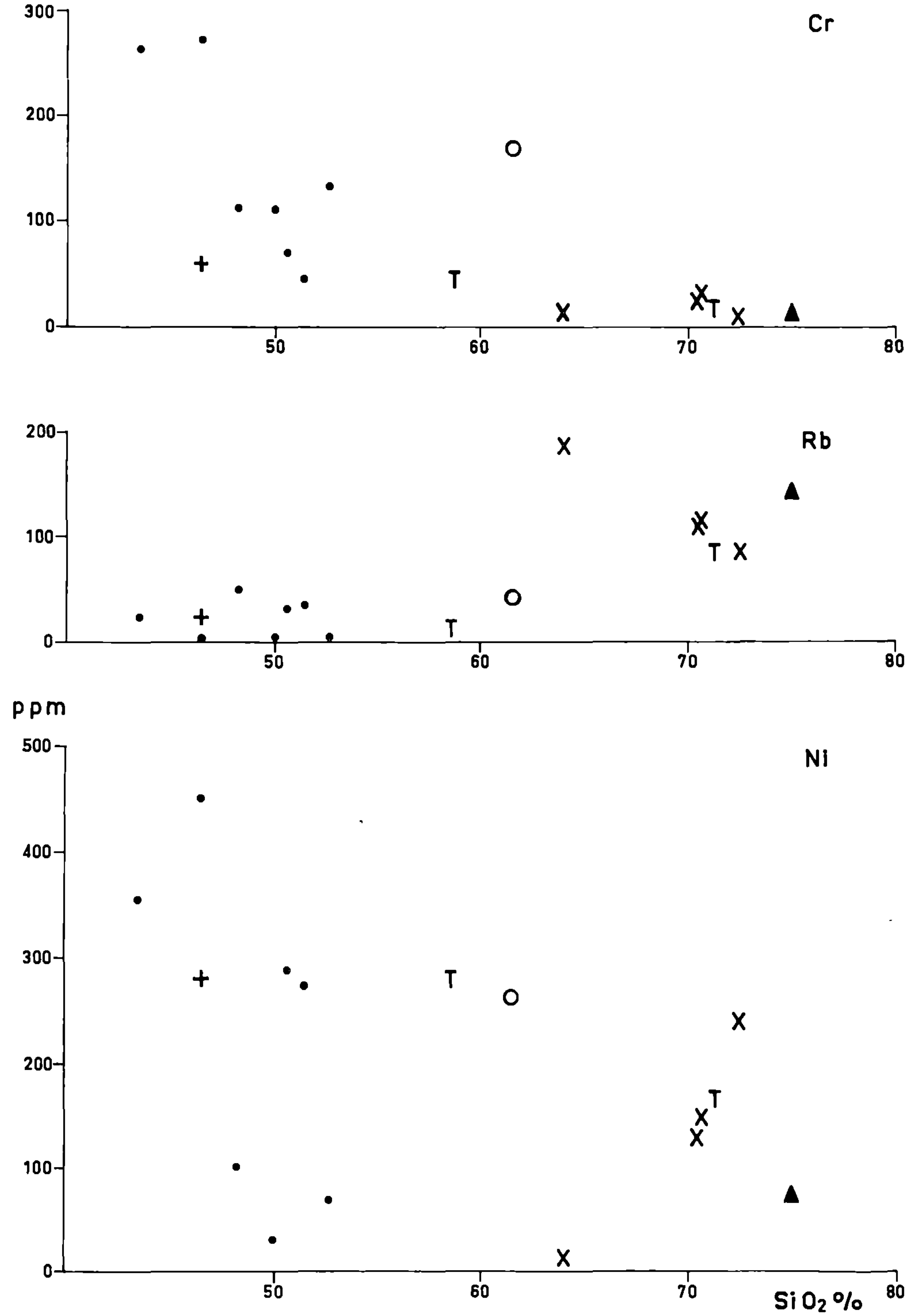
Trace Elements (in p.p.m.) in Tasmanian Keratophyric Rocks

	1	2	3	4	5	6	7	8
V	140	28	35	100	42	35	32	255
Cr	13	13	13	34	25	27	27	50
Ni	15	240	190	150	130	170	320	285
Cu	106	75	71		87			79
Zn	286	172	124	165	122	227	85	223
Rb	185	85	80	112	110	90	8	15
Sr	75	150	50	60	40	120	35	190
Mo	—	2	—	3	3	2	9	—
Ba	900	985	290	1050	4600	685	550	100
Pb	64	37	24		40			26

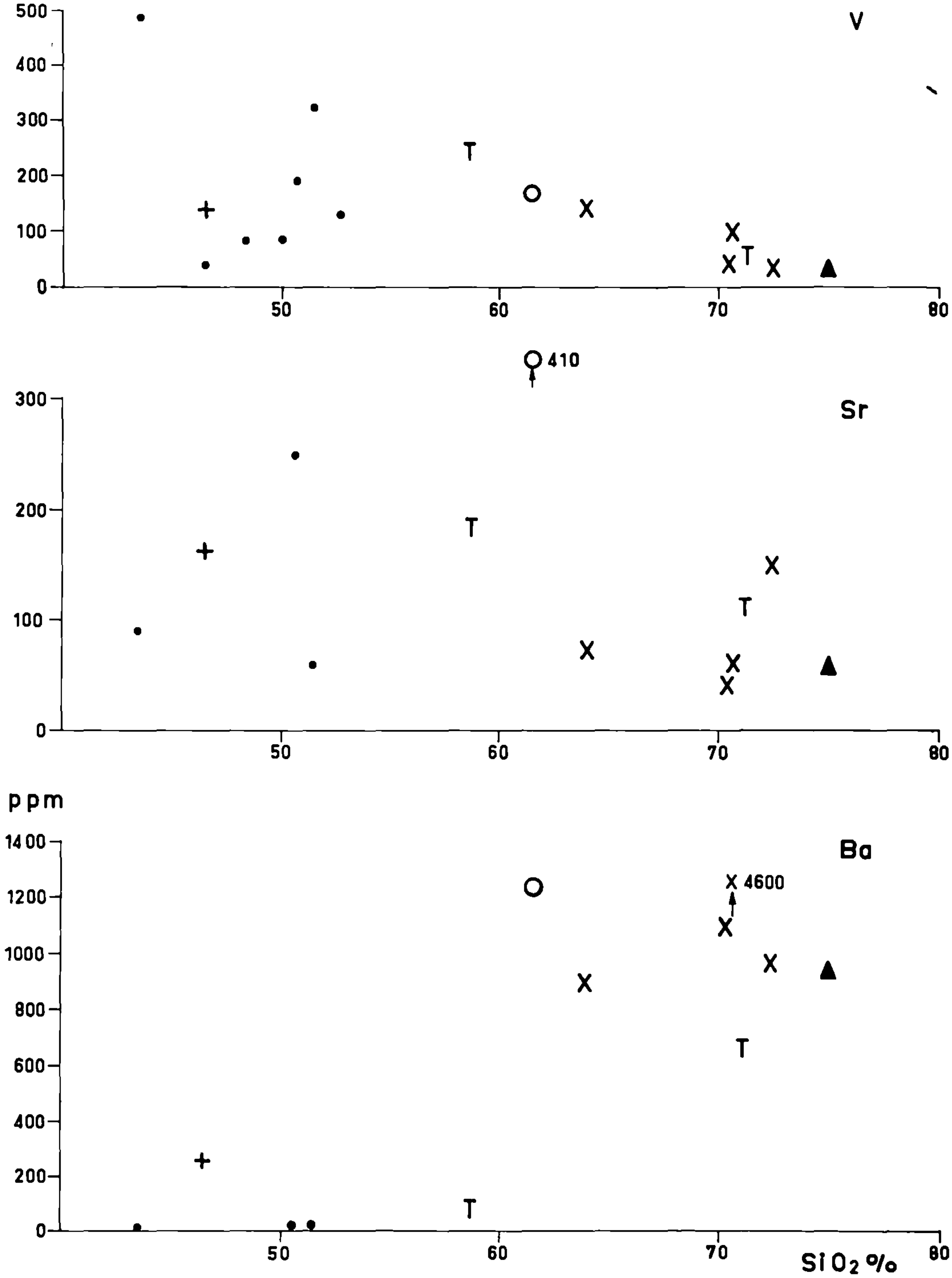
1. Quartz keratophyre, Lake Dora (31707A).
2. Quartz keratophyre, Comstock tram-line (31754).
3. Quartz keratophyre, one mile east of Rosebery (31760).
4. Quartz keratophyre, West Queen River (32538).
5. Darwin keratophyre, Mt. Sedgwick (31259).
6. Primrose Volcanics, north of Hercules (31789A).
7. Teepookana tuff (3177A).
8. Tuff, Lynch Creek.

Analyses by X-ray spectrograph (M. Solomon), except for Cu and Pb by Japan Analytical Chemistry Research Institute (spectrophotometer).

Fig. 21: (2 pages) Trace elements analyses of the Tasmanian spilites, keratophyres and related rocks, plotted against silica. Legend as for Fig. 15, analyses from Tables 6 and 13.



Cr
300
200
100
0
50
60
70
80
Rb
200
100
0
50
60
70
80
ppm
Ni
500
400
300
200
100
0
50
60
70
80
SiO2 %


V
Sr
Ba
ppm
410
4600
SiO2 %
0
100
200
300
400
500
1000
1200
1400
50
60
70
80

INTRUSIVE IGNEOUS ROCKS ASSOCIATED WTH THE VOLCANICS

Basic Intrusive Rocks

Intrusive Spilites

Cross-cutting "melaphyre" was first reported at Zeehan by Twelvetrees (1900a) who also described concordant melaphyres and tuffs within sandstones and shales. Twelvetrees and Ward (1910) confirmed this by reporting intrusive spilite near the Zeehan-Montana mine office. The discordant nature of some of the spilites of this neighbourhood are revealed on the map published by Blissett (1962) and reproduced as Fig. 22 and these spilites almost certainly fill feeder vents and dykes. The intrusive rocks and the flows are apparently identical though poor exposure and deep weathering have severely limited the range of the present investigation in which reliance has had to be placed on specimens gathered during mining and exploration work.

Similar feeders occur on King Island and form dykes and sills both in the spilites and in the underlying sediments. A typical sub-spilitic dyke crosses the dolomitic succession in Cumberland Creek (31814). It is about 10 feet wide and fine-grained. It consists of fresh augite and altered albite with very minor chlorite patches and a few crystals of ilmenite (?). The augite rarely forms good crystals and most individuals occur as broken crystals, fragments of crystals, and granules, of average size 0.05 mm. diameter. All stages in the cracking of crystals and the drawing apart of the fragments can be observed. Interstitial areas and gaps between crystal fragments are filled by highly altered albite laths and indefinite masses of calcite etc. There seems little doubt that feldspar crystallisation has followed crystallisation and jostling of the augite, the movement perhaps taking place during intrusion of the magma as a "crystal mush". Other intrusive rocks on King Island are similar but several are almost identical to massive lavas, particularly those cutting the volcanic succession.

Several more coarsely grained gabbroic rocks intrude sub-spilitic sediments from half to two miles south of Naracoopa, King Island, and will be described in the next section. In many cases it is extremely difficult to distinguish between spilitic flows and fine-grained gabbroic sills, particularly where field relations are not revealing. Thus near the Henty River Bridge (Zeehan-Queenstown road) several rocks assumed to be intrusives may in fact be rather coarse spilitic flows (e.g. 31734A, 31737A). These fine-grained intrusives are distinguished generally by close association with coarse-grained rocks, lack of amygdules and by the presence of ilmenite which is rare in known flows.

Gabbros

A number of small bodies of gabbroic rocks intrude the Cambrian sediments and volcanics. They are associated with the spilites and appear to be related genetically. They occur mainly within Crimson Creek rocks e.g. west of the Magnet Mine (Waratah district), in the Renison-Bell-Dundas area and south of Mt. Dundas. South of Naracoopa, King Island, gabbroic dykes and sills intrude sediments below the spilites. In the Henty River gabbroic sills intrude volcanics and sediments containing Middle Cambrian (i.e. Dundas Group) fossils.

The form of the gabbro masses is variable but generally sill-like; they are generally small bodies and the sills vary from several cm. to a few hundred metres thick. At Naracoopa they are sill-like and slightly transgressive, on the Henty River they appear to form lenticular sills; on Mt. Dundas, the mass is plug-like, and east of Renison Bell and west of Magnet they occur as both dykes and sills. Blissett (1962, p. 45) believes the Trial Harbour mass to be a sill of Cambrian age though it is possible that this gabbro is Devonian and related to the adjacent Heemskirk Granite.

The only ones already described are near the Heemskirk granite (Waterhouse, 1916; Blissett, 1962). The largest of these occurs between Zeehan and Trial Harbour and consists mainly of a granitic-textured aggregate of labradorite and hornblende with minor relict pyroxene. Waterhouse reports some orthoclase and albite. Magnetite, ilimenite and apatite are common accessory minerals and there are occasional grains of pyrrhotite.

Twin axis determinations on five crystals in 6898 and 4987 varied from An_{60} to An_{63} (labradorite) and plotted close to the low-temperature curve.

The hornblende prisms are as long as 8 cm. in coarse patches and from the writer's observations, are common hornblende with the following properties:

X, Y	:	very pale yellow
Z	:	pale green
α	:	1.635
γ	:	1.662
$\gamma - \alpha$	:	0.027
2 V	:	67 -
$Z \wedge C$	:	15°

Some of the crystals are partly replaced by a colourless to very pale green fibrous actinolite with relatively low birefringence, the Z axis of the actinolitic hornblende being at right angles to that in the earlier hornblende. Locally the gabbro is considerably altered, with complete clouding of the feldspars and alteration of amphibole to chlorite and actinolite (e.g. 4984).

An analysis of the gabbro is given in Table 14. A very similar rock outcrops north of Dundas where it forms a narrow dyke several hundred metres long. It consists (50G137) of 60% labradorite (fairly clouded) and 20% pale green to colourless hornblende, the remainder consisting of a felted aggregate of radiating or platy prehnite. A small mass of gabbro crops out north of the Trial Harbour mass and contains augite, partly altered to actinolite, as the principal ferromagnesian mineral. An analysis (Table 14, No. 1) indicates a considerable ilmenite content, and an unusually high water content, and it would appear that some of the feldspar is albite.

In all the other gabbros albite is an important constituent and they may be classed as albite gabbros or micrograbbros. They occur in the Henty River valley, south of Mt. Dundas, in the Ring River, west of the Magnet Mine and on King Island. The considerable difficulty in distinguishing the fine-grained albite gabbros from spilites has already been pointed out.

For a mile or so south of Naracoopa, King Island, the beach exposes a slightly transgressive sill, several hundred feet thick (31815, 31816). For the most part, the sill is composed of

Table 14: Gabbros

	1	2	3	4
SiO_2	46.60	49.14	46.44	46.65
TiO_2	2.69	0.59	0.47	1.00
Al_2O_3	14.30	13.81	19.71	14.57
Fe_2O_3	3.04	3.51	2.41	3.58
FeO	12.87	9.34	4.54	6.90
MgO	5.73	7.68	5.76	3.01
CaO	7.64	10.75	12.51	9.34
Na_2O	3.42	2.00	2.96	4.70
K_2O	0.36	0.87	0.96	0.48
H_2O+	3.20	2.24	3.73	2.80
H_2O-	0.40	0.03	0.39	0.27
MnO	0.08	0.19	0.11	0.18
P_2O_5	0.04	0.14	0.22	0.26
CO_2			0.07	6.44
S			0.02	0.01
Total	100.37	99.29	100.30	100.19

	5
V	120
Cr	60
Ni	285
Zn	136
Rb	25
Sr	165
Mo	-
Ba	260

Table 14: Gabbros (continued)

		1	2	3	4
qu					7.02
or		2.13	5.14	5.56	2.78
ab		28.94	16.92	25.15	39.30
an		22.61	26.14	37.53	5.84
cor					4.28
di		12.51	21.50		
hy	en	4.57	12.02	3.40	7.50
	fer	5.38	8.57	1.32	8.45
ol	forst	4.80	0.67	7.77	
	fay	6.23	0.53	3.37	
calc ortho				6.88	
mag		4.41	5.09	3.48	5.10
calc				0.20	14.00
ap		0.09	0.33	0.67	0.67
ilmen		5.11	1.12	0.91	1.98
Total		96.78	98.03	96.24	96.25

1. Gabbro, Heemskirk area (Blissett, 1962, p. 46).

2. Gabbro, Trial Harbour Road (Spry, 1962b).

3. Gabbro, Zeehan Road, ½ mile east of Henty River Bridge, specimen 31740A.

4. Fine grained gabbro, same locality, specimen 31799A.

Analyst: 1 and 2. Dept. of Mines, Tasm.
3 and 4. Japan Analytical Chemistry Research Institute.

5. Trace elements in p.p.m. in 31740 by X-ray spectrography (M. Solomon).

approximately equal quantities of augite granules and albite laths in a crudely granular texture. Again the augite granules appear to be disrupted portions of crystals and albite crystallisation seems later than the disruption. Small areas interstitial to the albite-augite framework are filled by chlorite, prehnite, calcite and possibly a little quartz and there are a few amygdules, up to 0.5 mm. diameter, filled with chlorite and calcite. 5-10% of the rock is composed of irregular masses of ilmenite. Parts of this sill are extensively altered and the primary texture may be completely destroyed e.g. 31826 consists of calcite, quartz, prehnite and leucoxene.

Between the Magnet Mine and the Whyte River there are several concordant gabbroic intrusions. Typical of the coarser gabbros is the body outcropping in the Corinna Road about ½ mile east of the Whyte River crossing (31865, 35N27). It is composed almost entirely of plagioclase, hornblende and chlorite (in order of abundance) in an even-grained micro-granitic texture, the average diameter being 0.5 mm. The feldspar is speckled with yellowish-brown chlorite and sericite (?) and many of the laths possess a thin clear rim in optical continuity with the cloudy core. The plagioclase is albite or albite-oligoclase. The hornblende is somewhat fibrous and ragged, and it appears to replace both augite (?)

and yellowish chlorite. Its optical properties are:

X	:	colourless
Y	:	yellowish green
Z	:	medium grass-green
α	:	1.635
γ	:	1.661
$\gamma - \alpha$	:	0.026
2 V	:	71 + ve

Magnetite and/or ilmenite and pyrrhotite occur as anhedral crystals. 35N27 (Mines Department) shows a tendency to ophitic texture.

East of this large body there are a number of thinner sills visible on the Corinna road (31862, 5124 35N14) and also at the Cleveland Mine (31864).

Sills less than a metre thick have been encountered in drill holes near the base of the Crimson Creek Argillite at Renison Bell. 31846 has intersertal texture, the largest crystals being stumpy laths of albite about 2 mm. long. Albite laths form 70% of the rock, the remainder consisting of crude, wispy hornblende crystals, 10-15% ilmenite crystals (largely leucoxene), actinolite, and several percent of pyrite and clear quartz. The hornblendes are partly converted to actinolite and there are irregular patches of actinolite needles and chlorite (?) that may have been horn-blende. Quartz occurs within albite crystals and in interstitial areas, particularly in association with quartz and actinolite needles.

A few hundred metres higher in the same formation, at the Ring River bridge, is a much thicker sill (31832) that consists of augite crystals up to 7 or 8 mm. diameter in ophitic texture with smaller albite laths; it contains several percent of leucoxene. The sill is patchily altered, both feldspar and augite being replaced by chlorite and prehnite. Veinlets, 2-3 mm. thick, of prehnite are common.

Gabbroic rocks crop out in the Henty River and on the Zeehan-Queenstown road, one mile east of the road bridge. They form small bodies of only a few hundred square yards, elongated parallel to the strike and probably sill-like. They (e.g. 31740A) consist of an aggregate (average grain about 0.75 mm. diameter) of very clouded albite, partly chloritised augite (2V averages 55 + ve) and ilmenite. Several of the feldspar laths have thin clear albite rims. Amygdule-like patches of prehnite are common and in parts the feldspars are converted to prehnite aggregates. These sills (?) are interleaved with keratophyric lavas and tuffs but no margins are visible. A chemical analysis of a prehnite-rich variety of 31740A is given in Table 14 (No. 3). Basic rocks in the vicinity show gabbroic, doleritic and porphyritic, almost basaltic texture and many are similar to spilites except that they do not contain amygdules. 31737A and 31799A have doleritic texture with albite laths up to 1 mm. long with interstitial granular augite, patches of chlorite and some late quartz, and several percent of magnetite (X-ray determination). The augite has the following properties:

α	:	1.679
γ	:	1.716
$\gamma - \alpha$	:	0.037
2 V	:	55 + average

It is therefore similar to the augite in the spilites. The chlorite is dark green, non-pleochroic, has low birefringence and β was determined as 1.640, rather higher than that in the spilites. A chemical analysis (Table 14, No. 4) indicates that some of the magnetite may be titaniferous.

Only two gabbroic bodies intrude the Mt. Read Volcanics. An augite-albite gabbro (31767A) crops out near the Strahan-Zeehan-Queenstown road intersection, near the western margin of the volcanic arc. The rock has a crude seriate texture dominated by euhedral augite crystals up to 2 x 1 cm. and as small as 0.2 mm. x 0.1 mm. The remainder (some 40%) of the rock is composed of an aggregate of albite laths up to 0.1 mm. long, and chlorite. Some of the augite crystals show slight marginal alteration to chlorite. A similar rock crops out west of the Murchison River bridge on the Murchison Highway, south of Tullah (30038).

An important feature of the West Coast gabbros is the presence of labradorite gabbros at much the same horizon as the spilites and albite gabbros. Despite the inference that the albite is derived from labradorite, no specimens have revealed evidence of this change, and the two types appear independent, though probably derived from the same magma.

Gabbros occur in rocks below spilites, of the same age as the spilites, and younger than the spilites. Their similarities of mineralogy, composition and distribution indicate a close genetic connection with spilites and they plot on the variation diagrams within the spilite fields (see Figs. 15-21).

Ultrabasic Rocks

Pyroxenite and serpentinite complexes make up the third unit of Steinmann's Trinity in Tasmania. No attempt has been made to study these rocks in detail and the following paragraphs are merely a summary of existing knowledge and reconnaissance work by the writer.

Ultrabasic rocks occur at Bald Hill and south of Magnet (west of Waratah), along the flanks of the Huskisson Syncline and extending into the Colebrook Hill-Renison Bell-Dundas area, near the Spero River and west of Macquarie Harbour. In the Colebrook Hill-Dundas area it has been fairly well established by Blissett (1962) that the ultrabasic rocks form thick sill-like bodies along the Crimson Creek-Dundas Group boundary. The main rock types are gabbro, norite, peridotite and pyroxenite (Ward, 1909) and Spry (1962b) considered the pyroxenite (enstatolite) to be the most abundant. Serpentinisation of the ultrabasics is widespread.

The only evidence for the age of these rocks is from Adamsfield where trilobites occur in a conglomerate containing serpentinite pebbles. The fossils date the weathering of the serpentinite as middle Upper Cambrian (Banks, 1962a). Early analyses of the ultrabasics appear to have been unreliable as they show high alumina contents and the only reliable analyses are those given by Taylor (1955).

There is no evidence to indicate whether the rocks are intrusive or extrusive - they may have poured out onto the sea floor prior to Dundas Group deposition or have been intruded during or at the close of Dundas deposition under a cover probably not exceeding

8,000 feet (2,600 metres). The ultrabasics clearly follow the onset of spilitic vulcanicity but their time relation to the gabbros is not clear.

Acid Intrusives

There are two granitic bodies within the Mt. Read Volcanic Arc: the Darwin and Murchison Granites. They are of Cambrian age and are closely related to the volcanics. The Darwin Granite has been briefly described by the writer in a previous thesis (1957) and the Murchison Granite by Brooks (1962). Further details are included in the following summary.

The Darwin Granite consists of two tabular sheets of different composition (pink and white granite), separated in places by volcanic rocks; in outcrop it measures about three miles by half mile (Fig. 3). The pink granite (31761, 32508) is the more abundant and forms the eastern strip; it consists of pink K-feldspar, colourless quartz and greenish plagioclase feldspar in typical granitic texture. A few ragged crystals of biotite and chlorite are seen in places and hematite is usually present. It proved impossible to obtain accurate lower refractive indices for the K-feldspar but the rather approximate values obtained ranged between 1.518 and 1.520 (31761, 32508). With 2V measurements ranging between 63°

and 70^{o} (Fig. 13) the K-feldspar is roughly similar to that in the potash keratophyres and may be termed orhtoclase cryptoperthite. X-ray photographs of crystals indicated the presence of albite in small quantity (using the (201) line intensity as a measure of concentration). No fine twinning is present but the crystals have indefinite extinction and contain slivers and blebs of albite. These are present as irregular patches up to 0.1 mm. across or in sub-parallel veinlets at an angle to the (001) and (010) cleavages; there are many cases in which the patches appear to have spread out from the veinlet margins. In some crystals albite patches occupy a greater area than the K-feldspar host. The veinlets, though crossing the cleavages, locally develop along cleavage to give a stepped or zig-zag appearance to the veinlets. The albite within the K-feldspars is everywhere in crystallographic continuity and almost in optical continuity with the K-feldspar host. The relationship between the two feldspars is not clear. Though the K-feldspar may have replaced the albite, representing a continuation of the processes observed in some of the lavas, the veinlets and the patches indicate the reverse relationship.

The large albite and K-feldspar crystals show "mutual boundary" textures without any sign of replacement of one or the other. These albite crystals appear to be of similar composition to those in the keratophyres (Figs. 9 and 10).

A modal analysis was made of the pink granite (32508) using a grid of 1mm. square and counting 800 points, thus giving a variance no greater than about five for the major constituents (see Hasofer, 1963 and Solomon, 1963). The results were as follows:

quartz	36%
K-feldspar	36
albite	26
chlorite	1
hematite	1
	100

The difference between this result and the norm (Table 15) is probably because a different specimen was used for the chemical analysis. The "granite" is in part closer to an adamellite, particularly in specimens in which the K-feldspar contains a high percentage of albite.

Trace element analyses of 31761 (Table 15) show a high nickel value (the analyses may not be reliable) and a rather high copper value, compared to other granites (see Rankama and Sahama, 1950).

Table 15: Cambrian Granitic Rocks

	1	2	3	4	5
SiO_2	59.56	73.70	74.96	71.9	76.92
TiO_2	0.74	0.23	0.13	n.dt.	0.19
Al_2O_3	14.74	13.00	13.55	14.7	14.07
Fe_2O_3	2.77	0.94	0.79	3.6	0.43
FeO	4.82	1.08	0.71		0.64
MgO	3.29	0.63	0.48	0.6	0.25
CaO	3.92	0.40	0.16	0.5	0.16
Na_2O	3.16	2.20	2.33	1.9	3.24
K_2O	4.82	6.40	5.57	4.7	2.61
H_2O+	2.06	0.75	0.86	0.6	1.84
H_2O-	0.10	0.25	0.16		nil
MnO	0.10		0.01		Tr.
P_2O_5	0.18		0.05		Tr.
CO_2			0.30	0.4	0.06
Total	100.26	99.58	100.06	98.9	100.41

	6	7
V	20	150
Cr	15	20
Ni	75	175
Cu	97	34
Zn	160	370
Rb	140	90
Sr	60	70
Mo	2	2
Ba	975	1,525
Pb	25	

Table 15: Cambrian Granitic Rocks (continued)

		1	2	3	5
qu		8.82	34.26	38.91	47.30
or		28.48	37.81	32.92	15.42
ab		26.74	18.34	19.72	27.42
an		11.80	1.95	0.47	0.41
cor			1.73	3.52	5.76
di		5.29			
hy	en	6.65	1.80	1.20	0.62
	fer	4.49	0.79	0.46	0.51
mag		4.02	1.39	1.15	0.62
calc					0.14
ap		0.42		0.12	
ilmen		1.41	0.46	0.25	0.36
Total		98.12	98.53	98.72	98.56

1. Murchison "Granite", Murchison River Gorge. Sample collected by C. Brooks. Analyst: Dept. of Mines, Tasm.

2. Murchison granite, Murchison River Gorge, specimen Analyst: Aus. Min. Develop. Labs.

3 and 4. Pink granite from Mt. Darwin, from Solomon (1960).

5. White granite from Mt. Darwin, from Solomon (1960).

6. Trace elements in p.p.m. in pink Darwin Granite (31761).

7. " " " " " " " Murchison Granite (31866).

Trace elements by X-ray spectrography (M. Solomon) except by Cu, Pb and Zn (Japan Analytical Chemistry Research Institute).

The <u>white granite</u> (32506) occurs on the west flank of the pink granite. It is almost pegmatitic and composed of quartz and altered albite in approximately equal quantity, and may be termed a granodiorite (see Table 15).

The Darwin Granite has no apophyses or associated dykes of pegmatite or aplite. Its contacts appear sharp and only in one place (west of Thompsons workings) has any hint of hornfelsed country rock been observed; Hills (1914a) did not see any baking of surrounding rocks. The granites intrude Darwin keratophyre and parallel the structural trends of the Jukesian Orogeny - they are thus more or less concordant. The composition of the pink granite is very similar to the potassic Darwin keratophyre and the white granite very similar to sodic quartz keratophyre (see Figs. 15-21). The writer has suggested elsewhere (1960) that the granite may have been intruded during the final phases of the volcanicity and probably lay less than 5,000 feet below the surface. Alternatively it may be an associate of the Jukesian Orogeny and intruded at a rather later stage into the folded (?) volcanics. It was clearly derived from the same magma as the volcanics.

Its pre-Ordovician age is obvious from the many boulders and pebbles of the granite (mainly pink) in the Jukes Conglomerate on South Mt. Darwin and Mt. Sorell; it is also considerably brecciated in places. Attempts to date the granite by K-Ar techniques have failed, probably due to the lack of ferromagnesian minerals and the alteration and deformation. Also contained in the Jukes Conglomerate are boulders and pebbles of hematite and hematised keratophyre. Hematite and magnetite veins up to several metres thick are common in the adjacent keratophyres and a few occur in the pink granite. Copper mineralization is associated with the iron ores, several veins containing chalcopyrite and pyrite. These Fe-Cu veins are essentially magmatic and related to both quartz keratophyres and the granite.

The Murchison Granite crops out in the Murchison River gorge and the outcrop covers about a half mile square (Fig. 23). It occurs within potassic and sodic quartz keratophyres and its margins are irregular and diffuse. It is vari able in colour and composition, varying from red to grey in colour and ranging between granite, adamellite, granodiorite and syenite, apparently without order and over short distances. This variation is enhanced by the presence of patches of potassic keratophyre within the granite. The gorge has onl y cut into the upper parts of the granite body and has penetrated no more than 60 m. below its top. The keratophyre lenses probably are small roof pendants.

The local variation in the granite probably helped to persuade Bradley (1954) that it was derived from granitisation of sedimentary rock. A large number of thin sections has been made available by the Hydro-Electric Commission, which has drilled and collected surface samples of the granite in several localities. The granite consists of quartz, albite, some oligoclase-andesine, K-feldspar, chlorite, biotite, hornblende, hematite and magnetite. These occur in crudely granitic texture which in many cases is marred by brecciation and twisting of crystals, particularly of the ferromagnesian minerals. In the more severely brecciated rocks, carbonate replacement is common.

The K-feldspar is much more homogeneous than in the Darwin Granite and only a few crystals show blebs and slivers of albite. Graphic intergrowth with quartz is relatively common and generally consists of blebs of quartz in the feldspar host. Several lower refractive index determinations gave approximately 1.520 and 2V measurements ranged from 50 to 61° (Fig. 13). These K-feldspars have a more limited 2V range, and possibly a higher percentage of albite molecule, than those in the Darwin Granite and the potassic keratophyres.

The albite is low-temperature, An_{0-10}, and shows albite, Carlsbad and pericline twinning (Figs. 9 and 10).

Concentrations of ferromagnesian minerals occur within the granite. In 30033, 10-20% of the rock consists of unaltered biotite and green hornblende plus magnetite, pyrite and alittle zircon, sphene and calcite. In this rock, the K-feldspar crystals enclose laths of altered albite. Pyrite is common in irregular patches of granite (e.g. 31864).

Analyses of the granite (Table 15) indicate a range from syenite to granite (300 30), and thin sections indicate a fairly wide range in K_2O:Na_2O ratio. A modal analysis (designed for a variance of 5) of 30028, an albite-rich phase, gave quartz: 39%; K-feldspar: 24%; albite: 39%; chlorite, hornblende and biotite: 16%; iron ores: 2%, indicating a composition between adamellite and granodiorite. Trace element analyses (Table 15 and Table 31) show high Ni and V and Zn figures (though Ni may not be reliable) and variable Cu and Pb values.

Despite careful search, no granite pebbles have been found in the Jukes Conglomerate outcropping at the west end of the gorge, 400 m. from the nearest granite outcrops. However, the granite has previously been regarded as Cambrian because it is locally deformed and is quite different from the Devonian-Carboniferous granites. These are relatively uniform in composition, tourmaline-rich, have contact metamorphic aureoles and associated aplitic and pegmatitic apophyses. K-Ar dating by I. McDougall (pers. comm.) indicates a Cambrian age for the Murchison Granite.

SEDIMENTS ASSOCIATED WITH THE VOLCANICS

Breccias, paraconglomerates, greywackes and mudstones occur relatively infrequently among the volcanic sequences but are important as indicators of the environment of deposition of the volcanic rocks.

Breccia-conglomerates very like the Jukes Conglomerate occur within the Mt. Read Volcanics e.g. north west of Comstock (Fig. 35) and east of Lake Dora (Figs. 24, 25). They consist mainly of boulders (up to 1 m. x 1 m.) and pebbles, angular to moderately rounded, of keratophyre and quartz keratophyre, and they are poorly sorted. Rare greywacke lenses show the bedding direction. The characteristic feature is the presence of well rounded pebbles up to 5 cm. diameter of grey vein-quartz or quartzite, presumably derived from a Precambrian source. The Jukes Conglomerate is distinguished merely by the fact that it conformably underlies the Owen conglomerate.

These breccias probably mark phases of sub-aerial erosion of the volcanics during which material from the Tyennan Geanticline was carried into the area. Similar breccia-conglomerates without rounded quartz pebbles occur sporadically through the volcanics and are impossible to distinguish from agglomerates and volcanic breccias.

An unusual rock is the "fuchsite breccia-conglomerate" that overlies (structurally) the Natone Volcanics west of Rosebery (X2580 Fig. 11). It is about 10 m. thick and is a poorly sorted rock with an open framework containing mainly subangular to rounded pebbles of pale grey chert and yellowish microcrystalline tuff, together with a few vein quartz and reddish limonitic pebbles. They tend to be elongated steeply in a well developed vertical cleavage and reach up to 5 cm. in length. The matrix consists of angular to subangular quartz, fragments of tuff and chert plus sericite (part chromian), carbonate, iron oxides, a little tourmaline, zircon, K-feldspar and albite. The tuff fragments are very similar to the Natone Creek tuffs and show variable sericitisation and carbonation. Carbonates appear to spread outside the fragment margins and to replace parts of the groundmass and at least some of it is post-cleavage. Campana and King (1963, p. 6) gave a partial analysis of the rock as follows: $CaCO_3$: 22.0%, $MgCO_3$: 12.9%, $FeCO_3$: 16.2%, insolubles: 47.5%, indicating a rather complex carbonate. No parts of the conglomerate examined by the writer ever reached anywhere near the 50% carbonate indicated by the analysis.

The conglomerate appears to be the result of rapid weathering of both volcanic rocks and chert and has undergone mild alteration. Carbonate growth and sericitisation is typical of alteration associated with mineralization at Rosebery and Hercules, though possibly some of the carbonate may have been syngenetic.

Sandstones in and below the volcanics have a varied composition and texture. Structurally below the Natone Volcanics are about 800 metres of quartz sandstones that extend north and south of Rosebery for several miles (Fig. 11). They are known as the Stitt Quartzite and consist essentially of thinly bedded micaceous sandstones alternating with finely laminated siltstones and containing fine conglomerate beds and lenses (31794A). Massive sandstone beds several metres thick form bold outcrops. Small-scale cross-bedding and intraformational crumpling are common in some horizons. Similar sandstones outcrop west of Williamsford and south of Mt. Dundas (in the Henty River) and probably represent the southerly continuation of the Stitt Quartzite.

A sandstone bed 30-60 m. thick (32559) forms the Miners Ridge south of Queenstown (Fig. 28); it is a quartz sandstone with subangular particles ranging from 0.02 to 0.3 mm. diameter. It contains several percent of micaceous material (sericite and some biotite) and a few altered feldspar fragments. It is tentatively correlated with the Stitt Quartzite. Micaceous sandstones occur as lenses to the west of Miners Ridge (e.g. 31234) and are interbedded with quartz keratophyres, spilites, tuffs, agglomerates and mudstones. Several hundred feet of similar sandstones (31798A) overlie tuffs on the Queenstown-Strahan road, 3 miles from Strahan.

Between Lake Beatrice and the Anthony River, along the east flank of the Mt. Read Volcanic Arc, are several hundred feet of grey quartz sandstones and conglomerates that underlie keratophyres near the Anthony River (Fig. 3) but are interbedded with volcanics near Lake Spicer (Fig. 24 and 25). The beds vary in thickness, reaching a maximum of 300 m. east of Lake Rolleston. The sandstones (31720A-22A) are micaceous and moderately well sorted, with an average grain diameter of about 0.03 mm. They are characterised by convolute folds and local areas of intense brecciation (slumping ?) marking a phase of deformation previously thought to be tectonic (Dr. B. Campana, pers. comm.). Examples of these structures are shown in Plate 44, Nos. 1 and 2.

Near Lake Rolleston north of Lake Dora, Owen-like siliceous conglomerates with rounded quartz and quartzite appear with the sandstones and Campana et al. (1958) regarded the sediments as Ordovician in age. However, the presence of these sandstones towards the base of the volcanics, and their westerly dip and the

Plate 44 Nos. 1 and 2:- Convolute (intraformational) crumpling in finegrained sandstones, north end of Lake Spicer.

presence of dolomite suggests a possible correlation with the sandstone-dolomite successions on the west side of the volcanic arc.

Sandstones (the Nubeena Quartzite and Slate) enclose the Montana Volcanics north and west of Zeehan. They consist essentially of quartz grains no greater than 1/16 mm. diameter in a matrix of sericite and clay (?) (Spry, 1964); from the mineralogy and two chemical analyses (p. 31) of rocks from Oonah Hill, Spry has termed them quartzose sub-greywackes. They are very similar to sub-greywackes underlying the Crimson Creek beds at Renison Bell.

Many of the greywackes described below may well be of tuffaceous or part tuffaceous origin. Those composed largely of quartz and rounded rock fragments are most probably greywackes, those with quartz and a high percentage of feldspar crystals may be partly tuffaceous. Rocks within this range are rare within the Mt. Read Volcanics but are common in the sediments and volcanics associated with spilites lying to the west of the Mt. Read Arc.

The rocks interbedded with the spilites and overlying the Precambrian sediments for three to four miles east of the Magnet Mine (Waratah) are typical (Fig. 7) (Groves and Solomon, 1964). Specimen 30659, from the old Magnet tram-line, consists mainly of isolated grains, up to 0.2 mm. diameter, of angular to subangular quartz and irregularly shaped feldspars in a ferruginous, magnetite-rich matrix. Though a few grains of fairly fresh albite are present, the majority of the feldspars appear to be microcline. Most of them show a very fine, spindle-like normal albite twinning and some show "grid iron" lamellae; they are optically negative and 2V measurements range from 79° to 87°. Several contain blebs and irregular patches of clear quartz and some crystals appear to be almost entirely replaced by quartz. The feldspars may well be of direct pyroclastic origin.

Other rocks from this area are similar (30655, 30657, 30658, 30660) but vary in quartz (up to 40%) and iron oxide content. Specimen 30657b is unusual in that the matrix has several percent of small (0.05 mm. diameter) hornblende crystals, apparently as a detrital component.

The abundance of microcline in a sequence containing albite lavas is unusual and has not been noticed elsewhere in Tasmania. It is suggested that the rocks are at least partly derived from volcanic explosions of acid and potassic lava which failed to form flows by reason of its high gas content. Quartz keratophyres have been found in this area but are albite rich.

Passing eastward along strike from the Magnet area and away from the spilites these tuffaceous (?) sandstones merge to sub-greywackes and greywackes with more marked sedimentary features. These are poorly sorted with an open framework and consist dominantly of angular quartz grains, several of which are elongate and exhibit undulose extinction. Other clastic grains include rounded albite, hornblende, augite, chlorite, magnetite and rare rock fragments. They range in grain size from 0.2 to 0.8 mm. diameter and generally the grains are not quite in contact. The matrix consist of chlorite, iron ores and small fragments of other minerals.

Similar tuffaceous sandstones (but with albite), greywackes and subgreywackes occur at a similar horizon east and west of Renison Bell, near Mt. Dundas, in the Henty River and west of Miners Ridge, generally associated with spilites, keratophyric tuffs and mudstones. They are correlated with the Crimson Creek Formation.

Sandstones rich in rock fragments are rare but do occur in the Crimson Creek beds. Specimen 50G136, from the N.E. Dundas tram, consists of moderately rounded particles of rock up to 1.5 mm. diameter in a fairly well sorted material with little matrix. The particles consist mainly of spilite, microcrystalline tuff and greywacke, together with quartz and some unknown fine-grained rocks. This rock is similar to the fuchsite breccia-conglomerate and is probably derived from weathering of underlying volcanic rocks.

<u>Siltstones</u> occur relatively rarely in the Mt. Read Volcanics (e.g. at Tullah, Red Hills and Comstock) and are interbedded with stratified tuffs. They are much more common west of the volcanic arc, interbedded with sandstones and volcanics. They are composed of detrital quartz grains scattered in a very fine-grained, sericitic matrix and generally show fine stratification (31789, 31762). On King Island, banded siltstones below the volcanics north of City of Melbourne Bay contain nodules of pyrite up to 3-4 cm. across that show crude concentric layering and are almost certainly of syngenetic origin.

Rather flattened ovoid bodies of pyrite with their long axes parallel to bedding occur in finely banded siltstones (31787-31790) at Red Hills that are interbedded with quartz keratophyres. These specimens were obtained from the core of three drill holes put down by the Electrolytic Zinc Co. and Rio Tinto Mining Co. in 1958.

A common feature of spilites is the occurrence of manganese in adjacent sediments e.g. Park, 1946 on Olympic Penninsula, and the manganes deposits of the Tamworth area in New South Wales (Segnit, 1962) which appear to be related to spilite occurrences. In Tasmania, manganese occurs in quantity in only one spilitic area, viz. Manganese Hill, immediately west of Zeehan. Here it occurs as pyrolusite with goethite and forms a large capping to Crimson Creek sediments and spilites carrying carbonates and silver-lead mineralization. Blissett (1962, p. 168) followed earlier workers in regarding the capping as derived from oxidation of manganiferous siderite and pyrite though he points out that some of the iron (and manganese ?) may be derived from the sediments. Drilling under the capping has proved that it does not extend far below the surface. The source of the manganese is not at all clear and though it may in part be derived from sediments leached by percolating acid solutions developed during sulphide oxidation, it may also be derived from manganiferous siderite of hydrothermal origin.

Another common associate of spilitic vulcanism in various parts of the world is chert, frequently of radiolarian origin (e. g. Cornwall, New South Wales). In Tasmania, several fine-grained siliceous rocks are known from the Crimson Creek formation, particularly near the base. For example, west of Waratah, the basal Crimson Creek sediments contain beds of pale grey chert up to 3 or 6 m. thick and possessing a characteristic sub-conchoidal fracture; they are exposed near the Magnet Mine, on the Corinna Road, and in the Cleveland Mine (31861, 31863). They are also reported to occur on the top of Mt. Cleveland (Reid, 1923). Fragments of similar material (31837) are common in the fuchsitic breccia west of Rosebery. They were probably derived from local chert lenses at this horizon such as the 5 m. bed (31845) encountered within tuff and brecciated rhyolites of the Mt. Read Volcanics during drilling at the Pinnacles Mine north of Rosebery. In thin section, all these rocks consist of a featureless microcrystalline quartz (?) aggregate.

At Renison Bell and Smithton, cherts occur below the Crimson Creek beds. The Renison Bell chert (the "Red Rock") immediately and conformably underlies mudstones and volcanics and might be regarded as the basal unit of the Crimson Creek formation. It has been mapped in some detail by the writer and P. A. Hill during surveys of the mine area in 1960 and 1961 (Fig. 26, 27).

This unit is extremely variable and two identical stratigraphic sections cannot be found. It is up to 35 m. thick near the mine but thins out in all directions - it has not been found outside the half mile square of the mine area. Rock types include pink or red chert, coarse-grained hematitic sandstone, mudstone, greywacke sandstone, coarse paraconglomerates with pebbles of chert and mudstone, and at least two lenticular beds of dolomite.

The cherts and hematitic sandstones are well exposed on and near the Rosebery-Queenstown road; in the cutting near the hotel, beds of pink chert up to 7 m. thick alternate with thin bands o f purplish grey mudstones and hematitic sandstones. Septarian nodules up to 15 cm. diameter are common in the thick chert beds and deformed chert nodules several cm. in diameter occur in mudstones a few hundred metres southeast of the cutting.

The reddish chert (32920) consists of a microcrystalline quartzose aggregate liberally speckled with hematite "dust". The chert nodules are laced with veinlets of fibrous "chert" that generally have a clear, coarsely crystalline quartz core.

The sandstones (e.g. 32921) consist of grains up to 2 mm. diameter, almost all of which consist of chert or very fine-grained quartz aggregate (up to 0.06 mm. diameter). These grains are rounded, in some cases almost bulbous and like the chert nodules in the mudstones, appear to have been plastic at the depositional stage. Adjacent grains tend to accommodate each other by moulding though some are separated by interstitial hematite. Very hematitic chert or hematite forms some grains. In a few samples, some of the chert grains have a ovoid shape and impart an oolitic appearance to the sandstone.

The chert (and probably the hematite) appears to have been a primary precipitate and to have been locally broken up or "necked out" to form sandstones and conglomeratic rocks.

The "Red Rock" marks the transition from Success Creek to Crimson Creek sedimentation and coarse conglomerates with quartzite, granite and volcanic rock boulders occur locally at its base, overlying dolomite. The hematitic chert sandstones are presumably derived from reworking of chert in shallow water. The presence of this chert at the level at which spilitic volcanicity commenced rather indicates that the silica and iron oxides were derived from volcanic exhalations.

Carey and Scott (1952) have described an oolitic chert from Smithton at a similar horizon to the "Red Rock". They regard the chert as silicified dolomite, the dolomite bed lying immediately below spilites, volcanic breccias and mudstones and overlying quartzites. The Smithton chert was first described by Nye, Finucance and Blake (1934) as a fine-grained silicified conglomerate interbedded with black chert and grey slates. It consists of grey and dark grey, ovoid to spherical (and a few angular) bodies of chert in a white siliceous matrix, the bodies varying between 0.25 and 0.5 cm. in diameter. Nye et al. believed the chert to be distinct from the dolomite but Carey and Scott suggested the chert is silicified dolomite, on the grounds that part of the dolomite is silicified. It seems fairly clear from the descriptions that the chert bands are

Fig. 22: Geological map of the Zeehan area to show the distribution of spilites compiled from maps by Blissett, Guilline, and Pitt.

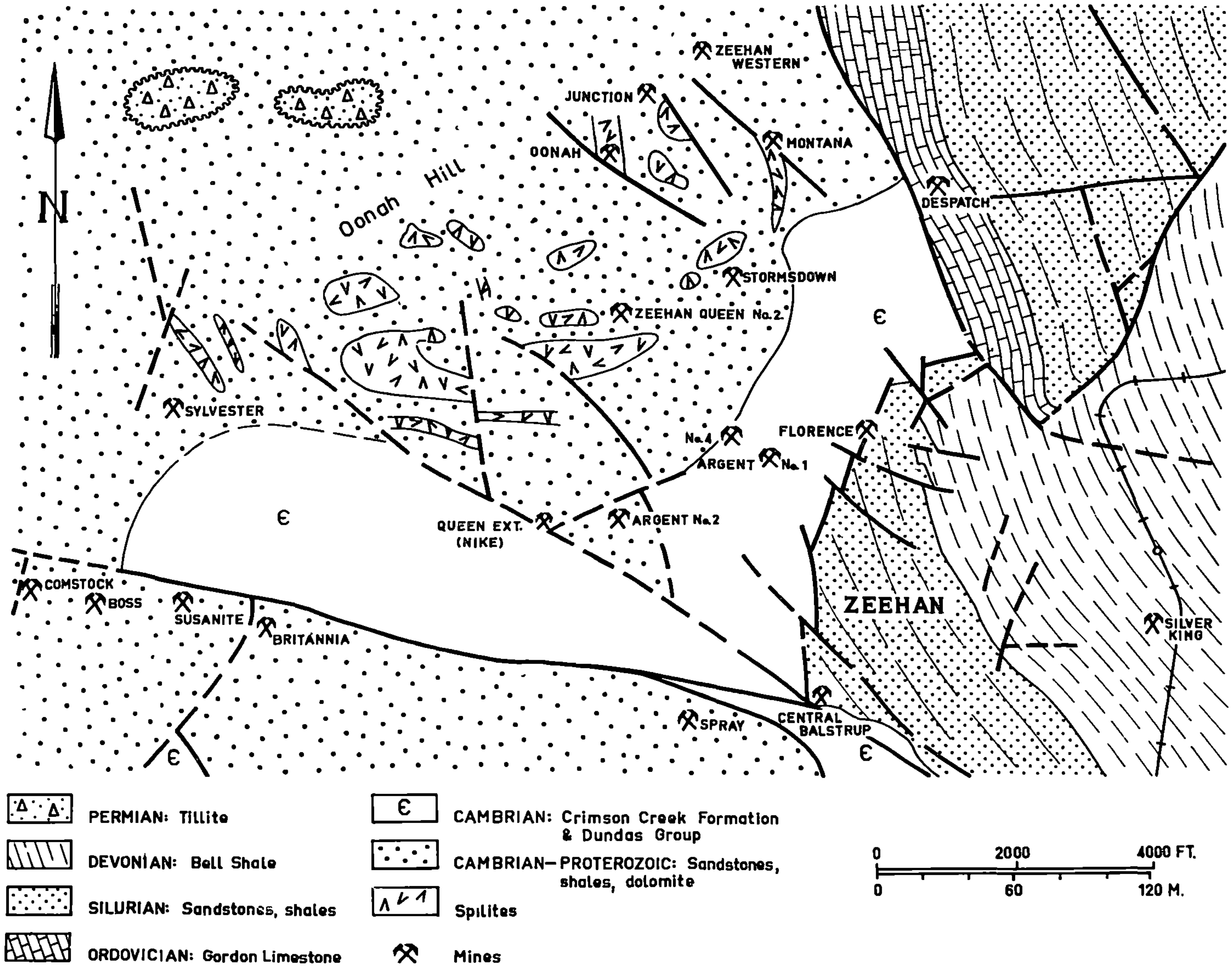
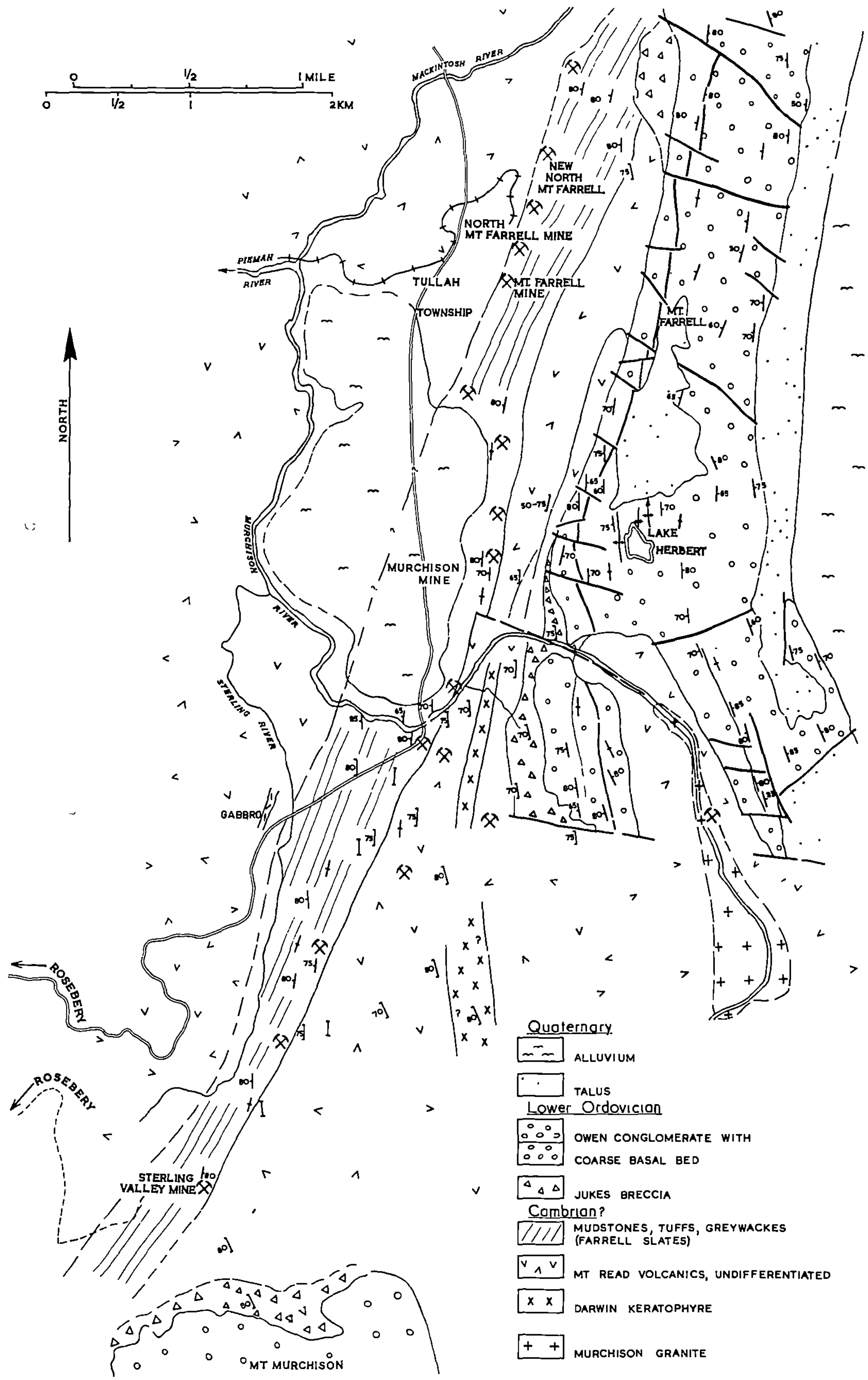
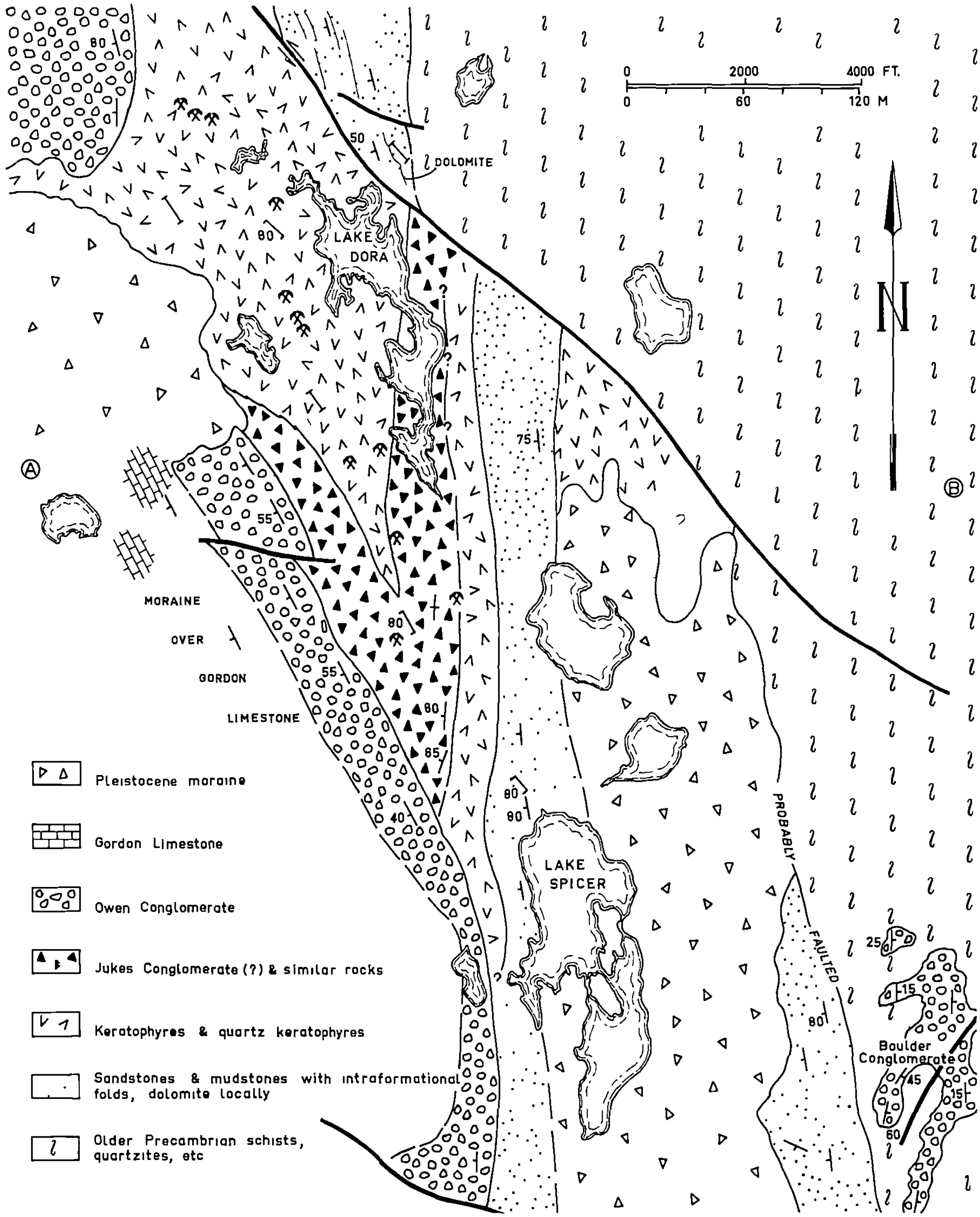

N
Oonah Hill
ZEEHAN WESTERN
JUNCTION
OONAH
MONTANA
DESPATCH
STORMSDOWN
ZEEHAN QUEEN No.2
SYLVESTER
No.4
ARGENT
No.1
FLORENCE
QUEEN EXT.
(NIKE)
ARGENT No.2
COMSTOCK
BOSS
SUSANITE
BRITANNIA
ZEEHAN
SILVER KING
SPRAY
CENTRAL
BALSTRUP
Є
PERMIAN: Tillite
DEVONIAN: Bell Shale
SILURIAN: Sandstones, shales
ORDOVICIAN: Gordon Limestone
CAMBRIAN: Crimson Creek Formation
& Dundas Group
CAMBRIAN–PROTEROZOIC: Sandstones,
shales, dolomite
Spilites
Mines
0
2000
4000 FT.
0
60
120 M.

Fig. 23: Geological map of the Mt. Farrel (Tullah) district. Geology by M. Solomon and C. Brooks.



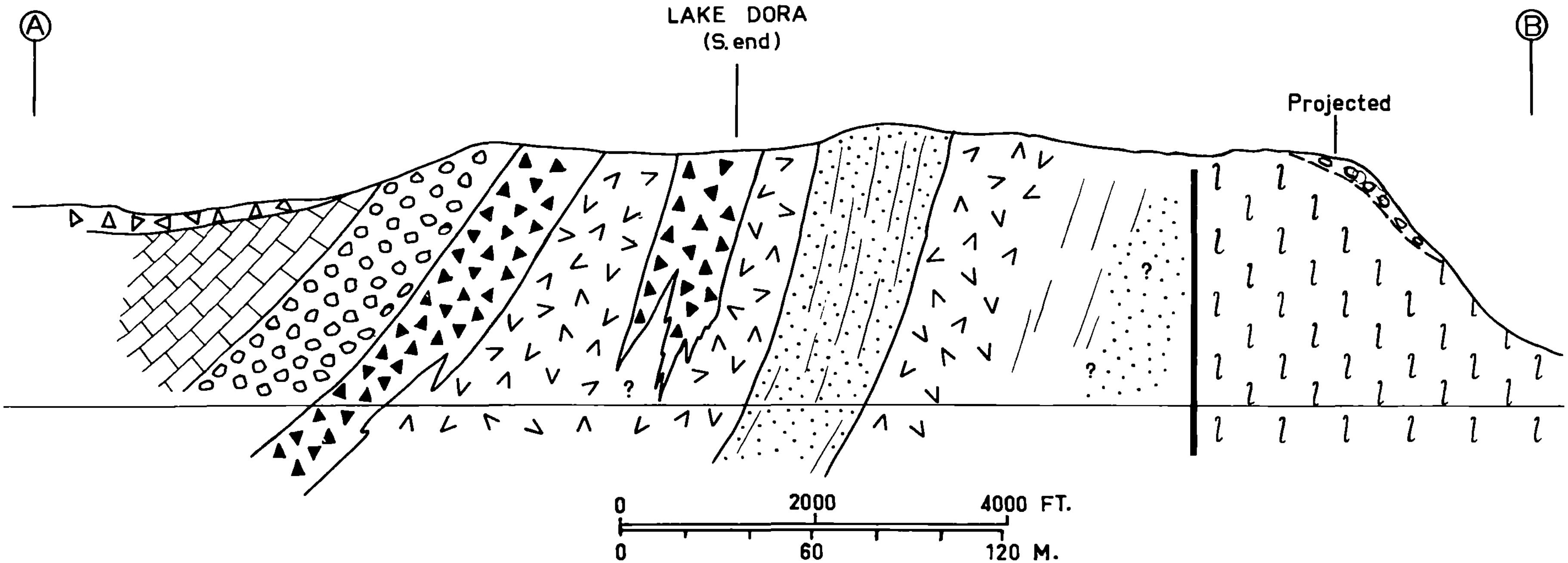
0 1/2 1 MILE
0 1/2 1 2KM
NORTH
MACKINTOSH RIVER
PIEMAN RIVER
NEW NORTH MT FARRELL
NORTH MT FARRELL MINE
TULLAH TOWNSHIP
MT. FARRELL MINE
MT. FARRELL
LAKE HERBERT
MURCHISON RIVER
MURCHISON MINE
STERLING RIVER
GABBRO
ROSEBERY
ROSEBERY
STERLING VALLEY MINE
MT MURCHISON
Quaternary
ALLUVIUM
TALUS
Lower Ordovician
OWEN CONGLOMERATE WITH COARSE BASAL BED
JUKES BRECCIA
Cambrian?
MUDSTONES, TUFFS, GREYWACKES (FARRELL SLATES)
MT READ VOLCANICS, UNDIFFERENTIATED
DARWIN KERATOPHYRE
MURCHISON GRANITE

Fig. 24: Geological map of the Lake Dora Area.



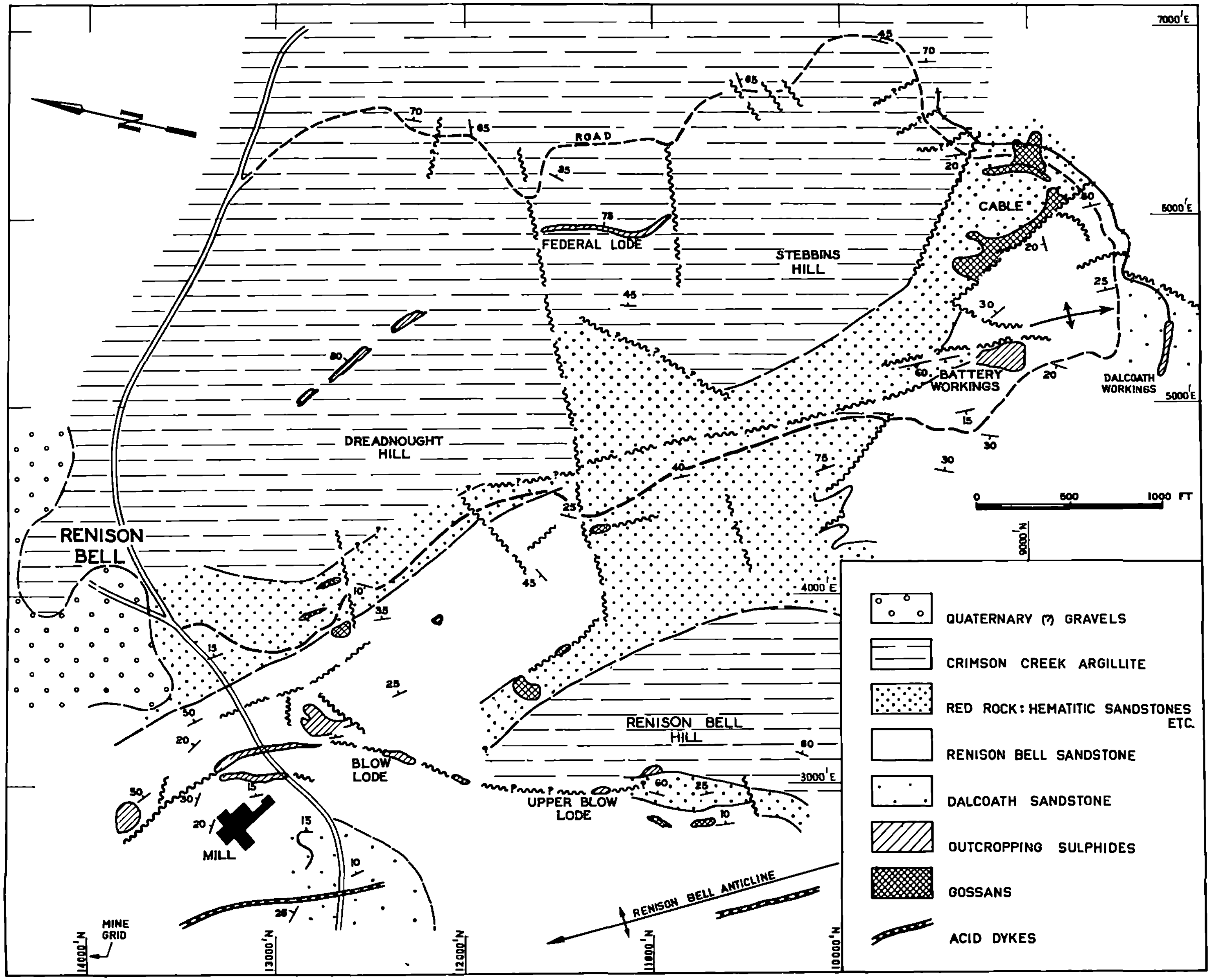
0
2000
4000 FT.
0
60
120 M
N
DOLOMITE
LAKE DORA
LAKE SPICER
MORAINE
OVER
GORDON
LIMESTONE
PROBABLY FAULTED
Boulder Conglomerate
Pleistocene moraine
Gordon Limestone
Owen Conglomerate
Jukes Conglomerate (?) & similar rocks
Keratophyres & quartz keratophyres
Sandstones & mudstones with intraformational folds, dolomite locally
Older Precambrian schists, quartzites, etc

Fig. 25: East-west cross-section through the southern end of Lake Dora (slightly generalised).



A
B
LAKE DORA
(S. end)
Projected
?
?
?
0
2000
4000 FT.
0
60
120 M.

Fig. 26: Geological map of the Renison Bell Mining Area. Geology by P. A. Hill and M. Solomon (1960-61).



N
ROAD
FEDERAL LODE
STEBBINS HILL
CABLE
BATTERY WORKINGS
DALCOATH WORKINGS
DREADNOUGHT HILL
RENISON BELL
RENISON BELL HILL
BLOW LODE
UPPER BLOW LODE
MILL
RENISON BELL ANTICLINE
MINE GRID
0 500 1000 FT
7000'E
6000'E
5000'E
4000'E
3000'E
14000'N
13000'N
12000'N
11000'N
10000'N
9000'N
QUATERNARY (?) GRAVELS
CRIMSON CREEK ARGILLITE
RED ROCK: HEMATITIC SANDSTONES ETC.
RENISON BELL SANDSTONE
DALCOATH SANDSTONE
OUTCROPPING SULPHIDES
GOSSANS
ACID DYKES

Fig. 27: Cross-section through Renison Bell Hill on mine gridline 1100 ft. North.

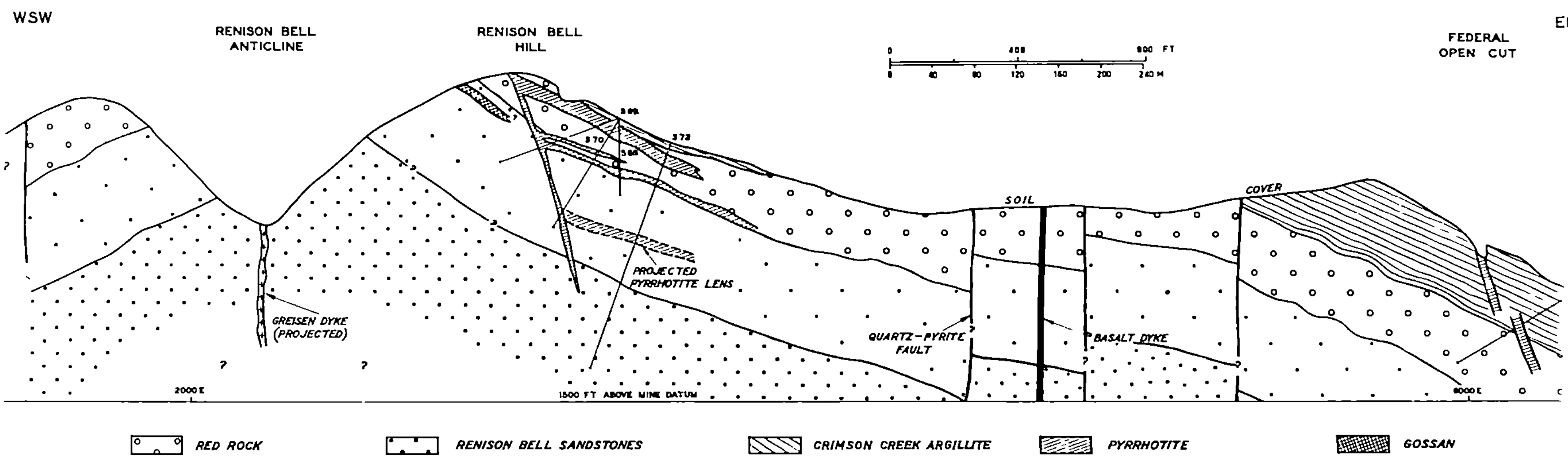
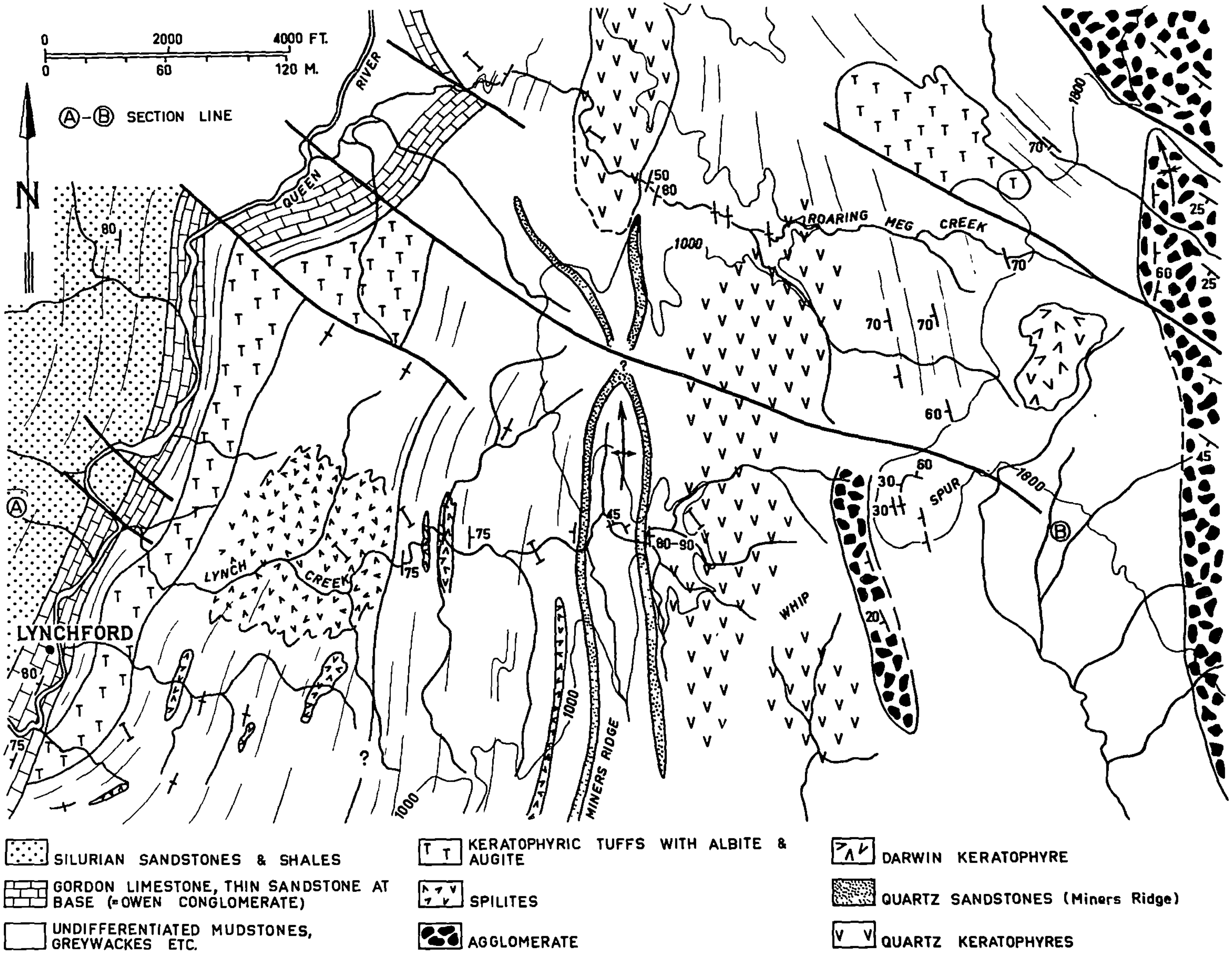
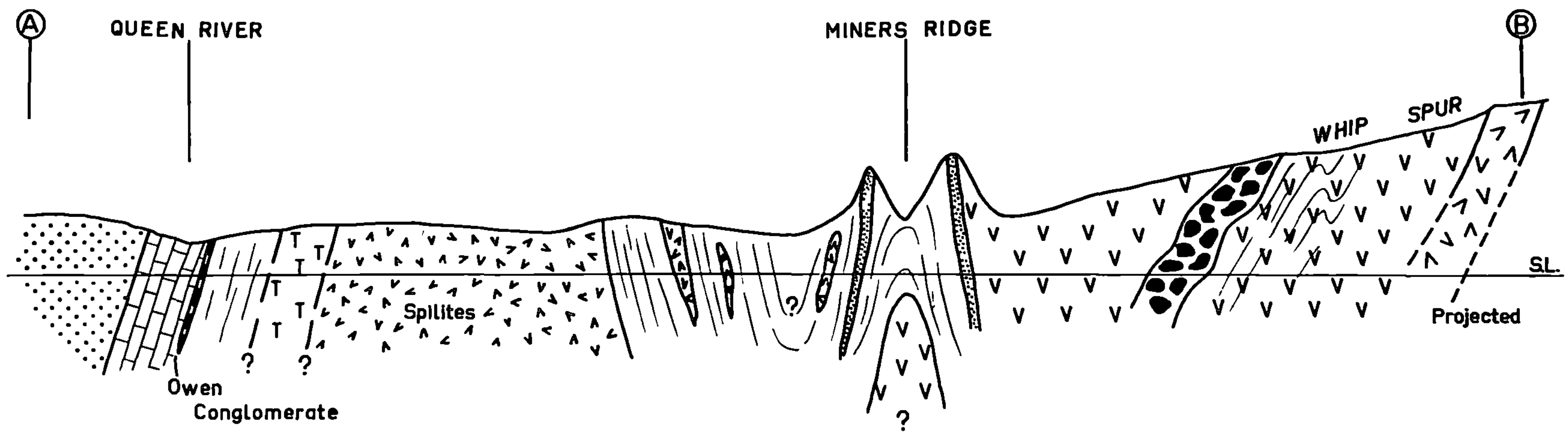

WSW
RENISON BELL ANTICLINE
RENISON BELL HILL
FEDERAL OPEN CUT
0 400 800 FT
0 40 80 120 160 200 240 M
370
372
PROJECTED PYRRHOTITE LENS
GREISEN DYKE (PROJECTED)
2000 E
1500 FT ABOVE MINE DATUM
SOIL
COVER
QUARTZ-PYRITE FAULT
BASALT DYKE
6000 E
RED ROCK
RENISON BELL SANDSTONES
CRIMSON CREEK ARGILLITE
PYRRHOTITE
GOSSAN

Fig. 28: Geological map of the Lynch Creek area, near Queenstown, to show the distribution of the spilites and keratophyres.



0 2000 4000 FT.
0 60 120 M.
N
Ⓐ-Ⓑ SECTION LINE
QUEEN RIVER
ROARING MEG CREEK
LYNCH CREEK
LYNCHFORD
MINERS RIDGE
WHIP SPUR
SILURIAN SANDSTONES & SHALES
GORDON LIMESTONE, THIN SANDSTONE AT BASE (=OWEN CONGLOMERATE)
UNDIFFERENTIATED MUDSTONES, GREYWACKES ETC.
KERATOPHYRIC TUFFS WITH ALBITE & AUGITE
SPILITES
AGGLOMERATE
DARWIN KERATOPHYRE
QUARTZ SANDSTONES (Miners Ridge)
QUARTZ KERATOPHYRES

Fig. 29: East-west cross-section along Lynch Creek.



A
QUEEN RIVER
MINERS RIDGE
B
WHIP SPUR
SL.
Spilites
Owen Conglomerate
Projected

primary units but that some dolomite has been silicified. The oolitic structure is probably derived from oolitic dolomite (see also Spry, 1964, p. 47).

To summarise, the sediments interbedded with the volcanics are important in that they throw some light on the conditions of deposition of the volcanics. The sediments are non-terrestrial and in part (probably entirely) marine and are derived from two main sources - a quartzose source (Precambrian rocks) and the local volcanic rocks. Chert beds close to volcanic horizons may have derived their silica from volcanic action.

STRATIGRAPHY AND DISTRIBUTION OF THE VOLCANICS

Two distinctive volcanic assemblages developed in the early Cambrian eugeosyncline of western Tasmania and they have fairly well-defined distributions. Along the eastern margin of the Cambrian basin, parallel to the Tyennan Geanticline, there grew a volcanic range of keratophyric nature, the Mt. Read Volcanic Arc, while to the west, in the deeper parts of the sedimentary basins, spilites and minor keratophyres were extruded and pyroclastic flows were probably swept in from the Mt. Read Arc.

Spilites north of Mt. Dundas and in Lynch Creek, close to the western margin of the Mt. Read Arc, are closely associated with keratophyric tuffs and tuffaceous sandstones.

Volcanic Rocks west of the Mt. Read Volcanic Arc

The spilites of King Island overlie mudstones, dolomite, and dolomite breccia (Fig. 4, 5). The breccia has been identified as tillite and correlated with the Sturt tillite of South Australia, of Cambrian age (Carey, 1947b). The assemblage of dolomites, quartz sandstones, and mudstones is one that occurs in several parts of Tasmania and in each case immediately underlying rocks that may be Lower or Lower Middle Cambrian. The dolomite-sandstone rocks may be early Cambrian or late Proterozoic.

In Cumberland Creek a spilite flow (31815) is interbedded with varve-like mudstones overlying tillite, and at City of Melbourne Bay Mr. H. Bartlett has pointed out a bed of spilitic tuff (31822) that lies within tillite.

At Smithton (Fig. 6) spilites overlie a succession containing dolomite, sandstone and chert and these beds have been correlated with the similar lithologies on King Island (Carey and Scott, 1952), and with other dolomite-bearing successions of the Younger Pre-cambrian (Spry, 1962a, p. 111). At Christmas Hills, west of of Smithton, Guilline (1959) discovered Upper Middle Cambrian fossils in a siltstone which he believed to be at the base of the volcanic-bearing succession. The writer has briefly examined the field relations in the area west of Smithton and it is clear that such a correlation cannot be supported as the gross structure of the Smithton-Christmas Hills area is unknown. From the meagre evidence available it is probable that the Christmas Hills rocks are considerably higher in the Cambrian than the spilites. At Penguin, pillow lavas overlie cherts, slates and greywackes that rest on Precambrian (?) rocks (Banks, 1956).

Near the Magnet Mine (Fig. 7) spilites occur near the base of a mudstone-greywacke assemblage correlated with the Crimson Creek Argillite (Groves and Solomon, 1964; Solomon, in press) and this assemblage overlies sandstones, shales and dolomite with apparent conformity. Quartz keratophyres occur in Crimson Creek sediments in the Corinna Road (639, 640). A very similar succession is to be found at Renison Bell (Hall and Solomon, 1962). Only thin gabbroic sills have so far been found there in the Crimson Creek beds, but Taylor (1954) reported a 9 m. "basalt" flow in the Crimson Creek Argillite in the Pieman River west of Renison Bell and spilitic tuffs occur near the base of the Crimson Creek Argillite (31845). It is possible that some of the gabbroic rocks are actually flows.

At Zeehan, the picture is not so clear. Spilitic flows, necks and dykes are found north and west of Zeehan (Fig. 22) in quartz sandstones, dark grey shales and thin dolomites, particularly in the region of the Western, Montana and Queen mines (King, 1961). The volcanics have been termed the Montana Melaphyre Volcanics (Hills and Carey, 1949) and the sediments the Nubeena Quartzite and Slate (Banks, 1956). The latter have been correlated with the Carbine Group (Blissett and Guilline, 1961) and included in the Oonah Quartzite and Slate of the Younger Precambrian (Spry, 1958, 1964; Blissett, 1962, p. 23). Spry (1964) suggested that the Montana Volcanics are discordant and are feeders to lavas in overlying Cambrian formations but there is little doubt that several of the spilites are concordant (Twelvetrees, 1900, p. 9) and no flows have been reported in the overlying formations.

The enclosing sediments are similar to the Oonah Quartzite and Slate except they contain bands of dolomite and the writer (in press) has correlated them with other dolomite-bearing successions immediately below the Cambrian (?) greywacke sequence. Twelvetrees and Ward (1910, p. 18) described "Keratophyric Tuffs and Breccias" in the mudstones overlying the spilite-sandstone succession at Zeehan, though from the writer's observations, the majority of fragmental rocks in the mudstones appear to be greywackes (e.g. 31847).

The Teepookana tuffs are exposed in an isolated fault block east of Strahan (Fig. 3) and overlie finely banded siltstones. The succession is almost certainly of Cambrian age and the siltstones may belong to the Crimson Creek Argillite.

The thick succession of spilitic volcanics in the High Rocky Point - WanderetRiver area overlies several hundred feet of sandstones, mudstones and tuffs with thin argillaceous limestones and calcareous shales (from reconnaissance mapping by the writer). Nye (in Finucane, 1932) mentioned small occurrences of quartz keratophyre associated with the spilites.

Amygdaloidal spilites (the Curtain Davis volcanics) occur north of Mt. Dundas in sediments and tuffs correlated by Banks (1956) with the Brewery Junction formation from the middle of the Dundas Group. Blisset (1962) tentatively accepted this correlation. The Curtain Davis volcanics are actually not within the type succession of the Dundas Group and the correlation just mentioned is based on superficial similarities of the enclosing sediments with Brewery Junction beds. A study of the geological map of the area shows that the volcanics are isolated from the Dundas Group by a series of major faults and are very close to the dolomite-sandstone rocks of Mt. Dundas (the Carbine Group), correlated by the writer with the Success Creek phase. The proximity of the volcanics to the base of the Crimson Creek Argillite suggests that they may be only a few hundred feet higher (stratigraphically) than the Zeehan spilites and quite possibly at a similar horizon to the "basalt" reported in the Pieman River by Taylor.

Associated with the Curtain Davis spilites are quartz-albite tuffaceous (?) sandstones or greywackes (50G146-158 and 169-172) probably derived from erosion of quartz keratophyres in the adjacent Mt. Read Arc.

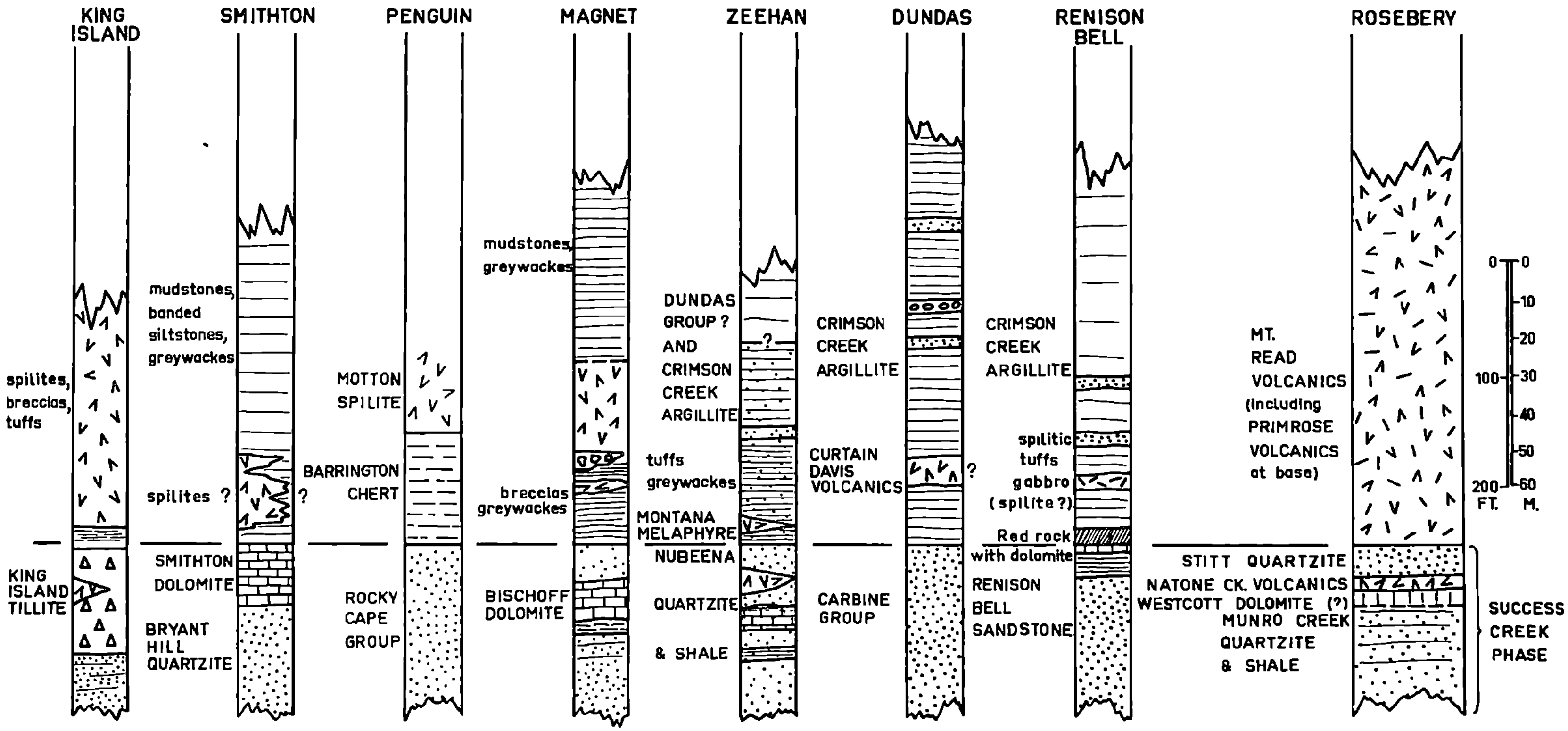
When Elliston (1954) described the Dundas Group he referred to major developments of volcanic rocks and in particular to basaltic flows in the Hodge, Brewery Junction, Comet and Climie formations. Banks (1956) suggested most of Elliston's pyroclastics were sediments but agreed that some tuff beds existed; these appear to be lenticular keratophyric tuffs but, like the Curtain Davis volcanics, do not occur in the type section and are correlated on lithological grounds. Blissett referred to only two thin bands of keratophyric tuff in the whole Dundas Group. The basic lavas mentioned by Elliston have not been recorded since, either by Blissett (1962) or during a recent study of the Dundas Group by Hills (1964). Banks (1956, 1962a), following Elliston on the distribution of lavas, believed the volcanics coincided with the siltstone phases of Dundas sedimentation but this idea has not been supported by other workers; vulcanism during deposition of the Dundas Group in the Mt. Dundas area appears to have been very limited.

A possible spilitic rock (or fine-grained gabbro ?) crops out on the Zeehan-Queenstown road, half a mile north of the Henty River. Its field relations are not clear but it could well be in the basal part of the Crimson Creek formation. In the Henty River (a quarter of a mile east of the bridge), the writer has found keratophyric tuffs (31800) interbedded with mudstones in which Blissett (1962, p. 40) has reported Middle Cambrain sponges. These tuffs, then, are of similar age to those reported in the type section of the Dundas Group.

In Lynch Creek, Queenstown, quartz keratophyric crystal tuffs occur interbedded with spilite flows and mudstones and the structural interpretation (Fig. 28, 29) indicates that this horizon is approximately equivalent to that of the quartz keratophyres to the east. The sandstone of Miners Ridge may be much the same age as the Stitt Quartzite.

In general, the Tasmanian spilites are not extensive, though up to several thousands of feet thick, and the individual occurrences appear to represent local centres of eruption. The diagram (Fig. 30)

Fig. 30: Summarised stratigraphic columns for various areas in West Tasmania. The datum line is the base of the volcanic-greywacke succession or the top of the sandstone-dolomite succession.


KING ISLAND
SMITHTON
PENGUIN
MAGNET
ZEEHAN
DUNDAS
RENISON BELL
ROSEBERY
spilites, breccias, tuffs
KING ISLAND TILLITE
mudstones, banded siltstones, greywackes
spilites ?
SMITHTON DOLOMITE
BRYANT HILL QUARTZITE
MOTTON SPILITE
BARRINGTON CHERT
ROCKY CAPE GROUP
mudstones, greywackes
tuffs greywackes
breccias greywackes
BISCHOFF DOLOMITE
DUNDAS GROUP ? AND CRIMSON CREEK ARGILLITE
MONTANA MELAPHYRE
NUBEENA
QUARTZITE & SHALE
CRIMSON CREEK ARGILLITE
CURTAIN DAVIS VOLCANICS
CARBINE GROUP
CRIMSON CREEK ARGILLITE
spilitic tuffs
gabbro (spilite ?)
Red rock with dolomite
RENISON BELL SANDSTONE
MT. READ VOLCANICS (including PRIMROSE VOLCANICS at base)
STITT QUARTZITE
NATONE CK. VOLCANICS
WESTCOTT DOLOMITE (?)
MUNRO CREEK QUARTZITE & SHALE
SUCCESS CREEK PHASE
0
100
200
FT.
0
10
20
30
40
50
60
M.

summarising the stratigraphic columns of most of these occurrences indicates a fairly consistent relationship between the Success Creek rocks and the volcanics. It would appear that spilitic eruptions heralded, or followed soon after, the onset of synorogenic, mudstone-greywacke sedimentation in the Dundas Trough. On the West Coast spilitic vulcanism appears to have practically ceased soon after the beginning of Dundas Group deposition though on the North Coast (near Ulverstone) spilites are recorded by Jennings and Burns (in Banks, 1962a) in the Upper Middle Cambrian. The keratophyric volcanics occur partly at the same horizon as the spilite flows and partly at considerably higher horizons, well up in the Dundas Group.

The Mt. Read Volcanic Arc

The age of these rocks has been the subject of controversy in recent years. Until a few years ago the volcanics had been regarded as equivalent to part of the Dundas Group (Carey, 1947a; 1953; Banks, 1956; and Solomon, 1960) but in 1958, Campana et al. indicated an age younger than the Dundas Group. Later (1960) they suggested the volcanics are older than this Group, though this was questioned by Banks and Solomon (1961) who considered that the facts brought forward by Campana et al. were not at variance with the admittedly tenuous hypothesis of contemporaneity with the Dundas Group. In this criticism the term Mt. Read Volcanics was first adopted.

In ~~1962 and~~ 1963, Campana and King summarised their previous views in a table that showed the Mt. Read Volcanics unconformably overlain by the Dundas Group and unconformably overlying the Success Creek sandstone-dolomite successions and also the Primrose Volcanics, which the writer has assumed to be part of the Mt. Read Volcanics. The critical evidence used by Campana and King to support their table appears to be as follows:-

(a) At Bulgobac the volcanics are overlain unconformably by Dundas Group sediments, and pebbles of volcanic rocks occur in the sediments.

(b) The volcanics are more sheared than the Dundas Group and Crimson Creek sediments.

(c) The volcanics overlie the sandstone-dolomite successions, referred to as the "basal Cambrian sediments".

Reasons for the unconformity at the base of the volcanics were not given and in fact Campana and King (in Banks 1962a, p. 131) stated that the Mt. Read Volcanics "rest on and appear to grade downwards to the early Cambrian succession discussed above....".

The writer has independently mapped what appear to be the critical areas for determing the age of the volcanics (i.e. the east and west flanks of the Mt. Read Volcanic Arc), and makes the following comments on Campana and King's conclusions:-

(a) Upper limit of the Volcanics

At Bulgobac the volcanics are overlain by mudstones, greywackes and tuffs (?) but there is no evidence of unconformity. The regional map, in fact, strongly indicates conformity (Fig. 3). The age of the overlying rocks is unknown (Banks and Solomon, 1961) but judging by the rock types they are likely to be equivalent to the Dundas Group or the Crimson Creek formation. They overlie the volcanics at Bulgobac but could be equivalent to volcanics that formed further east within the Arc; the tuffs (?) within the sediments indicate continuing volcanic activity. The rare pebbles of quartz keratophyre in the basal sediments are probably of very local derivation. The suggestion that vulcanicity continued well into Dundas Group time is supported by the occurrence of Dundas Group fossils in close field aseociation with the Mt. Read Volcanics near Valentines Peak. Pike (1964) following previous work in this area by A. Mackenzie (for the Electrolytic Zinc Co.) and the writer (for Rio Tinto

Mining Co.) has found that these fossils occur in mudstones within a few hundred yards of volcanics that are clearly part of the western margin of the Mt. Read Arc. The exact relationships of the rock groups are not clear but there is no evidence suggesting an unconformity or major faulting between them.

Further support for a Dundas Group age is provided by the occurrence of keratophyric tuffs in the Dundas Group near Dundas township (Blissett, 1962, p. 34) and in the Henty River.

Many of the tuff beds west of the Arc may be pyroclastic flows, derived from the Arc and contemporaneous with Mt. Read vulcanicity.

(b) Degree of Deformation

The volcanics and Cambrian sediments appear to have undergone very similar amounts of deformation and shear zones with intense deformation are found in both sediments and volcanics. The ubiquitous foliation of generally NNW to WNW trend appears to be of one age only and because it affects Silurian and Lower Devonian rocks, is assumed to be Tabberabberan.

(c) Base of the Volcanics

The relations between the "basal Cambrian sediments" and the volcanics is not clear though the weight of evidence favours the suggestion that the sediments (termed the Success Creek phase by Solomon, in press) immediately underlie the volcanics. The evidence for this is given below.

At Lake Dora (Figs. 24 and 25) the volcanics rest with apparent conformity on one or two hundred metres of sandstones and mudstones that in one place contains a siliceous dolomite (confirmed by X-ray identification). The succession east of the northern tip of Lake Dora is as follows:

- Quartz keratophyre -

120 m.:	Grey, poorly bedded medium-to fine-grained (average 0.1 mm. diameter) sandstone (31702A) with conglomeratic bands from a few cm. to 1 cm. thick, containing rounded white and pink quartz pebbles up to 2 cm. diameter.
30 m.:	Keratophyric tuff (31703A) with quartz and albite fragments in a chloritic matrix.
6 m.:	Contorted siliceous shales (31704A).
12 m.:	Pale grey fine-grained (average diameter 0.1 mm.) dolomite (31705A) containing 5-10% of rounded quartz grains.
3 m.:	Contorted siliceous shales with thin, fine-grained sandstone bands.

18 m.: gap - fault?

- Older Precambrian schists and quartzites -

On the west side of the Mt. Read Arc at about this latitude the exposures are not as good but in an east-flowing tributary of the Henty River, massive quartzite beds step down to the east in a series of asymmetric folds to disappear in the bed of the Henty River under mudstones and then massive keratophyric rocks (Fig. 3 point A).

Between Williamsford and the Silver Falls, the contact between Success Creek rocks and the volcanics is intermittently exposed. Much of this contact is faulted, judging by the discordant strikes on either side and the brecciation and crumpling of the sediments (e.g. 100 m. east of the Williamsford Hotel). Eight hundred metres north of Rosebery, the sediments dip easterly and appear, allowing for fault movement, to underlie the volcanics. Traverses along Higgins Creek and Lynch Creek (west of Pinnacles) by the writer (quoted in Campana and King, 1963, p. 13) reveal several thousand feet of mudstones, fine-grained greywackes and occasional conglomerates, dipping east at between 60° and 80° and showing only minor folding except near the volcanic boundary. Approaching Rosebery along the Pieman River (Fig. 11) or railway line from the north, the east dips steepen to vertical and half a mile north of the town, become westerly. From here to south of Rosebery the dominant dips in the sediments are westerly and appear to be overturned. Hall et al. (1953, p. 1148) considered these rocks were overturned ("proved by rock structures and by sedimentary characteristics") and Campana and King (1963, p. 7) state that overturning has been "confirmed by our investigations". However, no details are given and from conversations with these writers, it is clear that no definite proof of overturning has been found. Hills (1964) studying the structure between Renison Bell and Rosebery found two or three examples of small-scale cross-bedding (units 2 or 3 cm. thick) in a metre of sandstone in the Stitt River that very strongly indicates the west-dipping sediments are the right way up. Yet if this is extrapolated to indicate that all the west-dipping rocks are the right way up, then the east dips to the north must be overturned. The sediments west of Rosebery lie within an important fault zone striking NNE from Mt. Dundas and it is likely that they are severely faulted and/or folded. Thus without more evidence, which appears to be unavailable in present exposures, the problem must remain unsolved. However at present the regional evidence favours an interpretation placing the Success Creek rocks below the Primrose Volcanics.

The succession west of the Rosebery railway station was first described in detail by Finucane (1932), who termed it the Rosebery Series, and has since been checked by Campana and King (1963, p. 10) and the writer. From east to west it is as follows:-

approx. 1,300 m.:	Primrose Volcanics (regarded as intrusive by Finucane), containing lenses of finely banded mudstone.
640 m.:	Stitt Quartzite, a white or pale grey sandstone-quartzite, coarse-to fine-grained, thinly to thickly bedded and with rare conglomeratic lenses; shale partings are common. The ripple marks exhibited by Campana and King (1963) appear to be intersections of bedding and a fracture surface.
120 m.:	Natone Volcanics, mainly fine-grained tuffs.
90-120 m.:	Fuchsitic breccia-conglomerate, described on p. 188.
90 m.:	Westcott Dolomitic beds, reported by Finucane (1932) as purple slates and dolomitic siltstones. Quartzose greywackes are the only rocks discovered by the writer but carbonate-rich sediments.

are reported by Campana and King from this horizon both north and south of the Pieman.

+ 120 m.: Munro Creek Slate and Quartzite, alternating grey slates, carbonaceous slates and fine-grained sandstones.

The thick Stitt Quartzite does not continue as far north as the Chester mine area, either because it lenses out or because the beds become concealed by north-plunging folds.

The volcanic-Stitt Quartzite relationship is obscure and in most places faulted, and the writer knows of no evidence indicating unconformity as Campana and King suggest.

The "Rosebery Series" is similar lithologically to sandstone-dolomite successions at Renison Bell, Zeehan, Mt. Dundas, Mt. Bischoff, Smithton and King Island (see Fig. 30) and these are tentatively correlated under the term Success Creek phase. The Mt. Read Volcanics thus appear to immediately overlie the same sediments as the spilitic sequences and hence the lower part of the Mt. Read Volcanics is probably equivalent to the spilite flows.

The Lynch Creek (Queenstown) area provides further support for the contemporaneity of spilites and the lower Mt. Read Volcanics.

The distribution of the potassic volcanics ($K_2O > Na_2O$)

within the Mt. Read Arc may be of stratigraphic significance. The Primrose Volcanics on the west side of the Arc and the potassic rocks in the Anthony River-Lake Dora zone are clearly near the base of the Mt. Read Volcanics. The remaining zones of potassic rocks (Tullah-Red Hills, Madam Howard Plains, Mt. Sedgwick-Mt. Darwin) are probably in the basal parts of the volcanic pile but the structure in these areas is not sufficiently understood to be certain of the stratigraphy.

From the evidence available in the field, little better than a late Proterozoic to Cambrian age can be assigned to the Mt. Read Volcanics. However, a preliminary dating of the Mt. Sedgwick potassic quartz keratophyre (31257) by Mr. C. Brooks (using Rb/Sr methods) gives an age of 580 million years. This tentatively indicates an early Cambrian age for the lower part of the Mt. Read Volcanics.